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URANIUM EMPLACEMENT IN THE COLORADO PLATEAU

(Vice-Presidential Address, Section E, American Association for the Advancement of Science, Indianapolis Meeting, December 27, 1957)

BY PAUL F. KERR

ABSTRACT

Most Plateau deposits occur in sedimentary strata which range from Paleozoic to Tertiary, with major concentrations in Triassic and Jurassic formations. A number are primary and retain the characteristics of original deposition, but in many localities secondary action has obliterated the original features.

Four hypotheses of uranium emplacement on the Colorado Plateau justify consideration: (1) the hydrothermal theory, (2) the circulatory ground-water theory, (3) the ash-leach theory, and (4) the syngenetic theory. Observations indicate that (1) is most plausible in the light of recent evidence.

Additional wall-rock alteration data confirm the permeation of Plateau strata by hydrothermal solutions associated with uranium emplacement. Chrome-mica clay distributed along vertical fractures in sandstone, certain mica polymorphs associated with ore introduction, recrystallization of the clay-mineral constituents of sedimentary strata, and clay haloes associated with ore provide constantly accumulating criteria. Recently discovered alteration sequences in which dolomitization, argillization, and silicification accompany emplacement provide further evidence.

Mineral deposits of the Plateau yield 11 principal associated metallic elements: Se, Mo, As, Cu, K, V, Co, Cr, Pb, Zn, and S. In the interpretation of the temperatures of emplacement, the minerals which contain Mo, As, Cu, Co, and Cr particularly provide hydrothermal data. In these, distribution, association, occurrence in both hypogene and supergene forms, and general mineral history favor original precipitation from solution in the temperature range 100° to 350°C.

Organic materials heavily impregnated with uranium indicate uranium emplacement at temperatures essential to produce cracking of hydrocarbons. Hydrocarbon fragments from sediments entrapped in sulfide-uranium-bearing vein material yield clusters of spherical gas pockets, an apparent product of natural cracking. This would imply temperatures above 275°C. in places.

Uranium mineralization and alteration are associated with breccia pipes and collapse features which vertically pierce horizontal Plateau strata. Similar mineralization and alteration may be found near by and occupying channels of wide horizontal extent. Uranium-bearing diatremes, associated volcanic vents, and intrusives occur near by. The igneous features in places furnish regional patterns.

Acceptance of the data on absolute age, wall-rock alteration, mineral association, organic-ore emplacement, and vertical ore-bearing conduits with the time and temperature factors implied casts serious doubt on both the syngenetic and ash-leach theories. Extensive modification of the circulatory ground-water theory is also required to provide an original magmatic source for the uranium ions. The weight of evidence supports uranium introduction into Plateau strata by hydrothermal activity. Nevertheless, subsequent distribution by ground-water and superimposed supergene action must be recognized.

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INTRODUCTION

General Setting

The major uranium production of the United States is derived from deposits of the Plateau type in which ores are recovered from permeable strata. Similar conditions of occurrence prevail in portions of eight States from North Dakota on the northeast to Arizona on the southwest (Fig. 1). Most mining activity is concentrated in the region described by Fenneman (1917) as the Colorado Plateau (Fig. 2).

The origin of Plateau uranium deposits has long been a problem of geologic interest. Ores occur in a range of sedimentary strata and contain an extensive suite of uranium minerals. Both hypogene and supergene mineralization are involved. Ore-bearing areas are ordinarily horizontal or moderately inclined, but spread out laterally between enclosing strata. Faults,

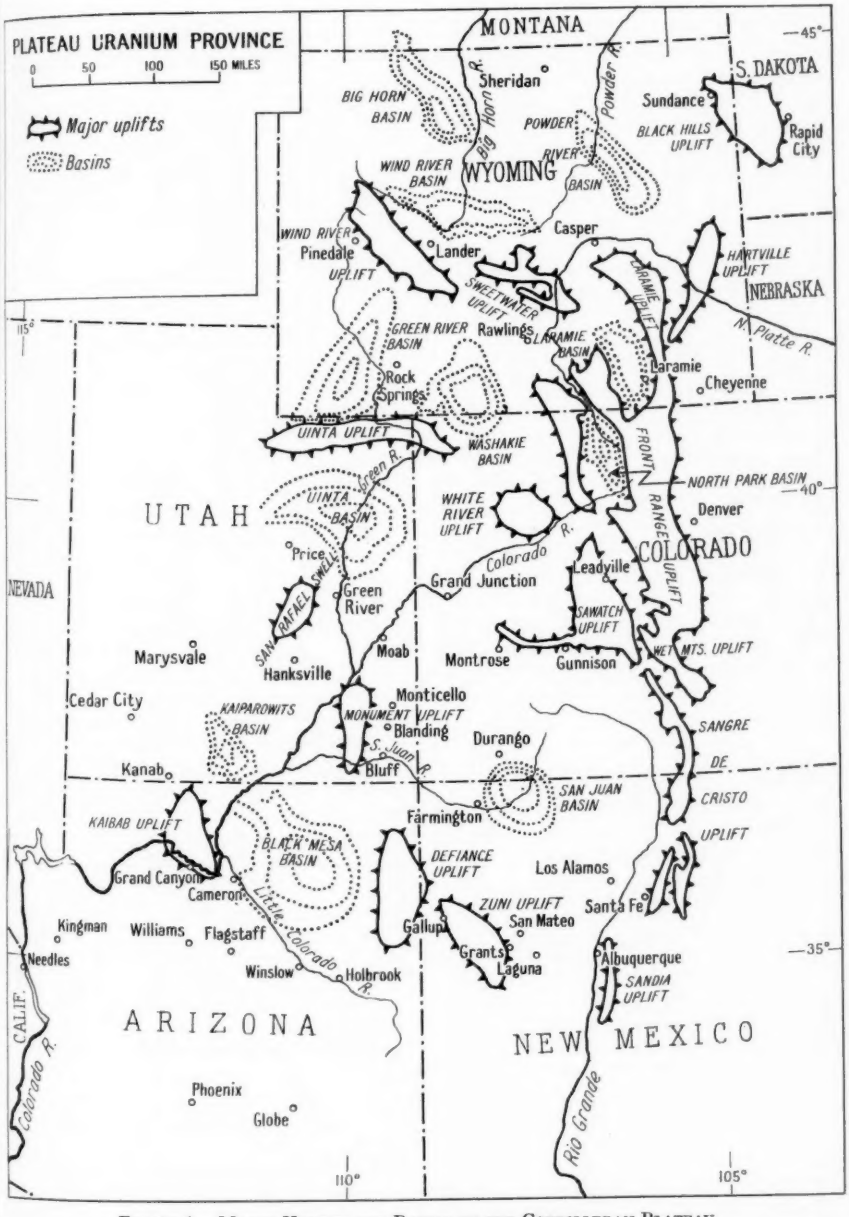

FIGURE 1.—MAJOR UPLIFTS AND BASINS OF THE CORDILLERAN PLATEAU

Uranium deposits of the Plateau type are associated with uplifts in a broad area which embraces portions of eight States.

FIGURE 2.—MAJOR URANIUM AREAS OF THE COLORADO PLATEAU

collapse features, diatremes, breccia pipes, and other vertical features which indicate possible conduits for mineralization are limited.

The writer examines the four most plausible hypotheses of uranium emplacement: the hydrothermal theory, the circulatory ground-water theory, the ash-leach theory, and the syngenetic theory. Evidence is then summarized and applied to the valid portions of each theory.

Hydrothermal Theory

The hydrothermal theory (Hess, 1922) envisages uranium deposition in sedimentary strata by precipitation from heated solutions. The primary source of both the uranium ions and the heat is attributed to igneous activity. Magmatic fluids similar to those observed around hot springs, thermal geysers, or fumaroles originate at depth. Such fluids represent the final gasp of igneous activity but are apt to mingle with ground water; dilution and drop in temperature may occur. These fluids provide the source of uranium ions.

The uranium-bearing fluids come in contact with a range of precipitants distributed through permeable strata. Reaction (Miller, 1955a) brings about the precipitation of such minerals as uraninite, coffinite, or urano-organic compounds. The precipitants may be plant detritus, petroliferous residues, argillic material, or their associated chemical constitutents.

Mineralization may be distributed horizontally along channels in sedimentary strata, it may impregnate porous strata, it may accumulate in fractured areas, or it may form at a fixed level determined by structure and ground water. Fractures in Plateau sediments provide conduits for the solutions, which migrate laterally for considerable distances. Supergene solutions above the water table may bring about pronounced changes in the primary mineralization. Many deposits originally hypogene are converted to accumulations of supergene minerals.

Circulatory Ground-Water Theory

The circulatory ground-water theory (Gruner, 1956) envisages uranium deposits emplaced in sedimentary strata by precipitation from normal ground-water solutions. Uranium ions are added to solutions by repeated circulation over exposed basement areas assumed to contain small amounts of uranium. Gruner has suggested that alkali and alkaline-earth bicarbonates would carry uranium in solution saturated with carbon dioxide. The temperatures are believed to be those of normal ground water. Not even a late-stage igneous influence is accepted.

The mechanism of precipitation may parallel that proposed for the hydrothermal theory. Lateral migration and supergene alteration would also occur.

This theory differs from the hydrothermal theory in two main essentials: (1) the uranium is derived by basement erosion, and (2) normal ground-water temperatures are assumed.

Ash-Leach Theory

Widespread deposits of volcanic ash exist in the Chinle, Morrison, and certain Cretaceous formations. It is assumed that the pyroclastic materials contain a small but consistent amount of uranium. This uranium is leached by meteoric water and precipitated at lower levels in favorable strata to form uranium deposits (Waters and Granger, 1953). These deposits are considered supergene. The extent to which this theory applies is a moot question. In Wyoming areas it has received strong support (Love, 1952). If applied generally, it precludes essential features of both the hydrothermal and circulatory ground-water theories.

Syngenetic Theory

The syngenetic theory (Wright, 1955) assumes that the uranium deposits were deposited contemporaneously with the enclosing sediments. Solutions low in uranium ions which accumulated during erosion of a distant basement land surface furnished repeated accretions of uranium for deposition in sediments. The mechanism of precipitation was essentially supergene, at low temperatures.

A few years ago this theory was widely accepted, prompted by the observed distribution in stratified deposits. If generally accepted, it would negate other theories.

Acknowledgments

Opinion has not yet crystallized concerning a number of factors in the interpretation of the emplacement of uranium, hence the benefit of contrary conclusions is readily available to one who ventures to write on this subject. Wholesome discussion during which well-considered differences of opinion have been expressed is cordially acknowledged.

The conclusions of this paper are derived

FIGURE 2.—MAJOR URANIUM AREAS OF THE COLORADO PLATEAU

from uranium studies which began in 1943. For a number of years this work has been made possible through a co-operative research agreement between the U. S. Atomic Energy Commission and Columbia University.

The writer is indebted to more than may be conveniently mentioned for discussion and numerous forms of co-operation. Assistance has been received from uranium operators in the field, staff and field officials of the U. S. Atomic Energy Commission, members of the U. S. Geological Survey, professors at several universities, and the mineralogical group at Columbia University. This aid is gratefully acknowledged.

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Interpretation of Uranium Emplacement

Available Criteria

In the interpretation of uranium emplacement it is essential to consider a wide range of criteria: (1) a comparison of the age of the enclosing strata with the age of the emplaced ores, (2) rock alteration associated with ore emplacement, (3) the nature of the primary ores, (4) the conditions under which minerals present are ordinarily precipitated, (5) the significance of the accompanying chemical elements, (6) laboratory experiments on the synthesis of uranium minerals, (7) the regional pattern of deposition, (8) structural factors involved in emplacement, and (9) the igneous centers of the area. Each item must be considered in terms of the postulated theories of uranium emplacement.

Absolute-Age Data

The criteria of absolute-age data are inconclusive. Few acceptable age determinations of Plateau deposits are on record. Numerous early determinations were made on oxidized material for which the values at best represent the time at which precipitation from supergene solutions took place. Absolute-age data on primary uraninite are also suspect unless supported by adequate mineralographic study. Microscopic amounts of galena contaminate many specimens. Unless the galena is separated and free galena is also present for an independent isotopic analysis, the significance of the lead-uranium ratio may be doubtful.

On the other hand, improved micro-analysis techniques coupled with careful field collection as well as microscopic control of laboratory separation and interpretation, at times yield absolute lead-uranium age determinations with plausible geologic interpretations (Table 1). However, few specimens meet these rigorous requirements.

Stieff, Stern, and Milkey (1953) give several age determinations for uraninite ore in Triassic Shinarump conglomerate with a range of 60-100 million years. Current data indicate that emplacement occurred after deposition of the enclosing sediments and over a considerable time range. The existence of a range in times of emplacement favors the hydrothermal theory, but the extent of the range is still too inadequately defined to be considered definite.

Recent determinations from Lisbon Valley specimens indicate that the radiogenic lead content of galena and uraninite from the same deposit are essentially the same. Where such natural radiogenic contamination exists, the migration of solutions bearing radiogenic lead appears to vitiate the age data.

Alteration

General discussion.—Plateau strata with abundant colorful red, brown, and green beds are bleached to white or light gray in many places near fault lines, uranium ore bodies, collapse features, and breccia pipes. This bleaching frequently signifies alteration and appears to accompany uranium mineralization as a prelude to precipitation. Much of the bleaching resembles the alteration along veins and faults found in igneous rocks as at Marysvale, Utah (Kerr *et al.*, 1957), where hydrothermal solutions have developed argillic alteration in the wall rock.

Bleaching associated with uranium mineralization is shown in places on the San Rafael Swell (Fig. 3). The effect may occur along fault

lines, in collapse areas, or in strata associated with ore (Fig. 4).

In places, solutions rising along structural openings have removed calcium and magnesium from carbonate rocks at a lower level and deposited large masses of dolomite (Fig. 5) above. The complete replacement of extensive quartz masses which has resulted is believed to require hydrothermal conditions. The marginal development of clay minerals which occasionally accompanies the dolomitization places both in the category of hydrothermal products.

Sideritization, alunitization, and silicification are more localized. Siderite probably corresponds to dolomite in origin. Silicification is less definite (Pl. 1, fig. 1) but in places is closely related to dolomitization and argillic alteration. Alunite may be due to the action of acid solutions on aluminum-bearing minerals. In places, however, hydrothermal activity may be involved.

Argillic alteration.—Argillic alteration yields bleached zones along faults (Figs. 6, 7) accompanying ore zones and within mineralized channels. It occurs both near ore and considerable distances from ore. Among the most significant argillic minerals found in Plateau sediments are recrystallized kaolinite, certain polymorphs of mica, and chromium-bearing mica clay (Kerr and Hamilton, 1958).

Close to the ore at Temple Mountain, Utah, veinlets of recrystallized kaolinite in sandstone become more prevalent. They also appear with greater frequency within and bordering collapse structures. Mica clay (illite) gains prominence, and the $2M_1$ polymorph becomes more abundant. This is accompanied by an increase in aggregate birefringence as well as in particle size.

Syntheses of mica polymorphs by Yoder and Eugster (1955) have shown that the transition to the $2M_1$ polymorph appears to occur between 200° C. and 350° C. at a pressure of 15,000 pounds per square inch. Although kaolinite may crystallize at lower temperatures, as shown by its occurrence as a weathering product, the higher temperatures are consistent with its formation in many places.

One of the most persistent argillic alteration effects is a mottled red and white pattern which occurs in limited areas in the siltstone strata of the Chinle formation and is described locally as "pinto rock." It is found above and below the ore zone at Temple Mountain, in altered areas elsewhere in the San Rafael Swell, at the base of the Shinarump in the Cameron area, in exposures along the Colorado River near Castle Valley, and has been found in cores from a wild-cat well drilled along the Salt Valley anticline. It is prominent in cores from the Temple

TABLE 1.—URANIUM-LEAD AGE DATA
(Primary Uranium Ore, Colorado Plateau)*

Locality	Geologic Occurrence	Laboratory†	Age: M.Y.**
Cane Creek, Utah, West of Moab	Vein in altered wall rock, Cutler formation. Permian	LGO-K 153	53
Happy Jack Mine, Southeast of Hite, Utah	Channel deposit Shinarump (Chinle formation) Triassic	SS-64	65
Temple Mountain, Utah, Southwest of Green River	Urano-organic ore in wall rock of Moss Back (Chinle formation) Triassic	LGO-K 154 SS-182	84 80
Orphan Mine, South side of Grand Canyon, Arizona	Breccia pipe penetrating base of Coconino formation. Permian and older wall rock	LGO-K 168	87
Lisbon Valley, Utah, Mi Vida Mine, South of La Sal	Channel deposit in Moss Back (Chinle formation). Just above unconformity at top of Cutler formation. Permian	LGO-K 169 LGO-K 152	185 190

*Ages selected are those for which both microscopic and mass spectrometric data have been available.
†LGO = Lamont Geological Observatory, courtesy of Mr. D. S. Miller and Dr. J. L. Kulp; SS = Stieff, Stern, and Milkey (1954, p. 264).
**Ages (206/238) may be subject to as much as ±10 per cent error; M. Y. = millions of years (approximate).

Mountain collapse. Large Calex cores of mottled Chinle siltstone are a prominent feature of the mining district at Temple Mountain (Pl. 1, fig. 2).

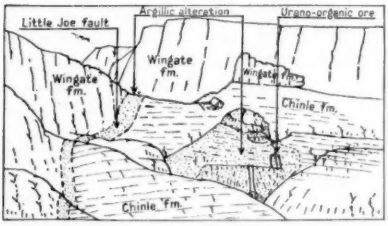

FIGURE 3.—URANIUM MINERALIZATION AND ALTERATION ACCOMPANYING FAULTING

Little Joe prospect on the west side of the San Rafael Swell, Utah, looking west. Urano-organic ore occurs in veinlets and nodules along a minor fault in an altered area (right center of sketch). An alteration zone also follows the Little Joe fault at the left. The Wingate formation shown in the background is about 400 feet thick.

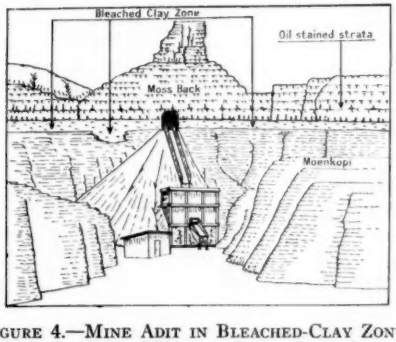

FIGURE 4.—MINE ADIT IN BLEACHED-CLAY ZONE

Dirty Devil No. 3 Mine on the southeast slope of Tomsich Mountain, San Rafael Swell, Utah. Uranium ore is found in Moss Back strata above the bleached clay zone. Moss Back strata away from the ore zone are oil-stained. The bleached-clay zone marks the base of the Chinle formation. Below the unconformity is the Moenkopi formation.

Lack of continuity, occurrence near faults, concentration near ore bodies, and argillic recrystallization indicate that the mottled effect grades into more intense alteration accompanying ore deposition. The mottled zones grade in places into the more intense kaolinization found beneath ore bodies.

A greenish chromium-bearing mica clay occurs in and near collapse structures at Temple Mountain along fractures and is disseminated among sand grains in the Wingate sandstone. It appears to represent an alteration effect associated with the emplacement of uranium mineralization.

Carbonate emplacement.—An impressive area of carbonate emplacement is exposed south of the main collapse on Temple Mountain, Utah (Pl. 2, fig. 1). Here a mass of Wingate sandstone about 200 feet thick and 700 feet long is almost completely replaced by dolomite. Quartz has been removed, and numerous solution pockets are lined with cockscomb dolomite crystals. Large dolomite masses appear on the north slope of Temple Mountain (Fig. 5). Nodular dolomitic replacement of Chinle strata may be observed on the west side of the San Rafael Swell (Figs. 5, 6; Pl. 2, fig. 2). Rounded carbonate aggregates range from a fraction of an inch to 12 inches in diameter. Many larger nodules contain cavities lined with dolomite crystals. The nodules are also surrounded with a halo of greenish-white clay in a marked color contrast to the prevailing red of the Chinle shale. Examples of dolomitic replacement are

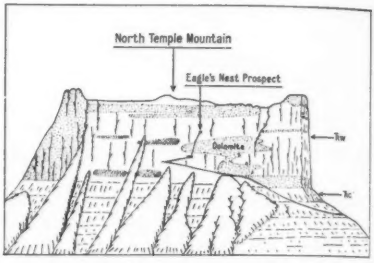

FIGURE 5.—DOLOMITE MASSES REPLACING QUARTZ SANDSTONE

Looking south at the north escarpment of Temple Mountain, San Rafael Swell, Utah. The vertical cliff at the right rises about 400 feet. Small amounts of urano-organic ore have been mined in the adits shown in the view. Dolomitic replacement occurs mainly in the Wingate formation, a massive quartzose sandstone.

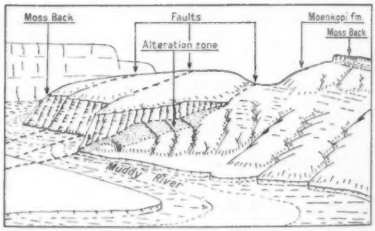

FIGURE 6.—ARGILLIC ALTERATION ALONG A FAULT

Looking westward along the Muddy River at west margin of San Rafael Swell. An alteration zone may be observed in Moenkopi strata bordering Moss Back (Chinle). Uranium occurrences of Tomsich Mountain lie to the east along the same structural zone.

shown in strata beneath the Moss Back on the west side of the San Rafael Swell at the Green Dragon (Figs. 10, 11), Conrad (Fig. 9), and Sulphur Springs Canyon (Fig. 8). Masses of

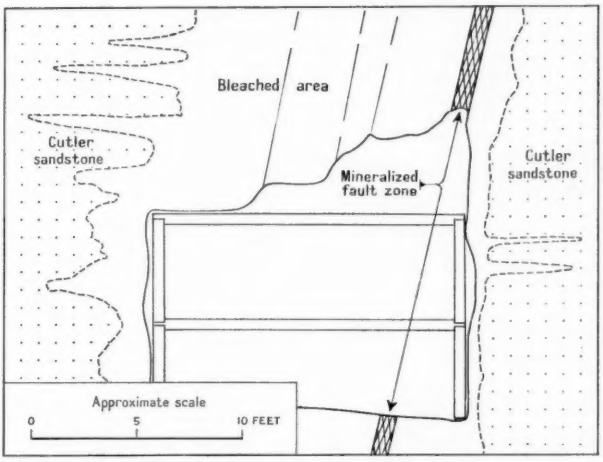

FIGURE 7.—ARGILLIC ALTERATION BORDER OF URANIUM VEIN
A small vein at the Honey Bee Prospect, Cane Creek, Utah, with bordering alteration in the Cutler sandstone.

dolomite also replace Wingate sandstone at the Rajah Mine east of the La Sal Mountains.

Large masses of siderite have formed in the sandstones of the uppermost Chinle and the Wingate at Temple Mountain. These are less extensive than the dolomite replacements and appear more closely related to ore zones.

At Temple Mountain, hydrothermal solutions sufficiently acid to dissolve carbonate appear to have ascended through the Kaibab limestone. The Kaibab was almost completely dissolved with a resulting collapse of overlying strata. Solutions ascended through the brecciated rock into the porous Wingate sandstone where dolomite was deposited. Greater alkalinity may have been achieved through contact with potassium-bearing clay in intervening strata (Fig. 12).

Silicification.—The occurrence of silica layers below dolomite in a similar relationship in four separated localities on the western side of the San Rafael Swell is not a coincidence. Sediments several feet thick intervene between the dolomite zone and the silica zone in each instance, yet conditions of precipitation existed. Clay alteration accompanied both dolomite deposition and silicification. Calcium and magnesium were probably dissolved from near by sediments, and silica was released. Carbonate above and silica below were probably precipitated because of slight differences in pH. The formation of silica deposits by the action of hydrothermal solutions on silty carbonate rocks has been noted in hot-spring areas, as at Golconda, Nevada (Kerr, 1940).

At the Green Dragon No. 3 Mine on the San Rafael Swell near the base of the Chinle formation, and separated from the Moenkopi by a thin gray clay is a silicified zone up to 5 feet thick (Fig. 11); the zone is mineralized with uranium.

The silica layer lies just below a channel cut into a "pinto zone" and filled with Moss Back strata. The silica layer was ore bearing, but apparently the Moss Back itself did not yield ore, although the structure of the deposit is partly concealed, and the continuity of the ore has not been completely tested.

The channel lies on the south flank of an anticlinal spur which departs from the San Rafael Swell northwest of Tomsich Mountain and extends westward toward the Last Chance Mine about 10 miles away. In the intervening area is a considerable zone of igneous activity where dikes and sills are widely distributed. No direct connection between the igneous activity and uranium deposition has been established. However, it seems reasonable to attribute the collapse features of the Swell to an aftermath of hydrothermal solutions, and a concentration of silicification of the type observed at the

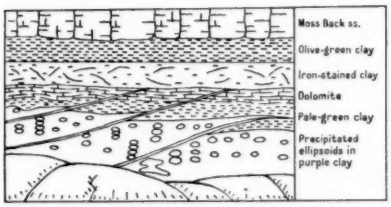

FIGURE 8.—DOLOMITE NODULES AT BASE OF CHINLE FORMATION

An exposure of 20 feet of strata on the south side of Sulphur Springs Canyon (0.37 mile from Red Canyon), west side of the San Rafael Swell, Utah. Pale-green clay fills fractures and forms haloes around the dolomite nodules. A large dolomite–clay aggregate several feet across is shown in the lower part of the outcrop. Many of the dolomite ellipsoids are about a foot in length and show vertical alignment parallel to minor axes.

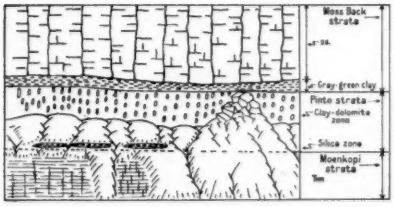

FIGURE 9.—DOLOMITE NODULES AND SILICA ZONE AT BASE OF CHINLE FORMATION

Near Conrad Prospect on the west side of the San Rafael Swell, Utah. The gray-green clay is 2–3 feet thick. About 6–8 feet of dolomite-bearing strata are exposed. The silica zone is 12–18 inches thick.

Both dolomite and chalcedony are associated with haloes of greenish white argillic alteration. Unaltered shale is brown to reddish purple.

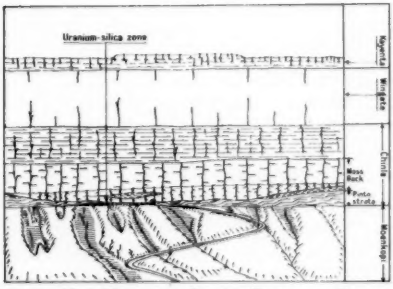

FIGURE 10.—URANIUM–SILICA ZONE AT BASE OF CHINLE FORMATION

Green Dragon No. 3 adit on east slope of an escarpment, northwest of Tomsich Mountain, about 1 mile north of the Muddy River, southwestern portion of the San Rafael Swell, Utah. The silicified zone is about 5 feet thick in places and extends for several hundred feet. Uranium minerals occur in a solid mass of quartz and chalcedony.

Green Dragon Mine would fit such an interpretation.

Less extensive silicification occurs in the ore of the Lucky Strike Mine which is located about opposite the north flank of the anticlinal spur mentioned above. The ore body of the Lucky Strike Mine lies near the base of the Moss Back in a channel. An adit cut below the ore shows extensive alteration which has even invaded the Moenkopi at this point. Both dolomitization and silicification may be observed in "pinto" strata on the canyon wall opposite the mine. Silicification also occurs below the dolomite near the Conrad Mine.

The interpretation of the silica–dolomite relationship from the standpoint of ore deposition is not entirely clear. However, four occurrences are known on the western side of the Swell where ore is near by, and this silica-dolomite relationship has been observed.

Silicification is a prominent feature near the base of the Chinle formation in the Cameron area. Residual angular fragments of red, slightly radioactive chert are strewn over wide areas. These occur particularly south of the Little Colorado River on the southern fringe of the uranium area.

Alunite.—In and around the area of uranium mineralization at Marysvale are large deposits of alunite. These bodies are remarkably deficient in radioactivity in contrast even to surrounding unaltered volcanic rocks. The alunite deposits are limited to the tuffs and intercalated flows. Some concentrations of uranium mineralization have been found in the flows; in the tuffs, however, there are only a few sporadic occurrences of weak mineralization along faults. The alunite bodies, as pointed out by Willard and Proctor (1946), may be enclosed in an envelope of rock rich in kaolinite.

In the Tushar Range alunite bodies occur in veins and replacement masses in volcanic rocks high on mountain ridges. Exposed in canyons below are sediments believed to be Carmel, Navajo, Kaibab, Coconino, and Mississippian. Thin, quartz-bearing veins are mineralized with copper, lead, zinc, and some pitchblende. Alunite bodies have not been found in the sediments, and, conversely, in this area the metallic mineralization has not been found in the overlying volcanic rocks.

The relationships suggest that the same solutions which deposited alunite in the tuff series formed metal deposits in the sediments below.

The abundance of hydrothermal alteration

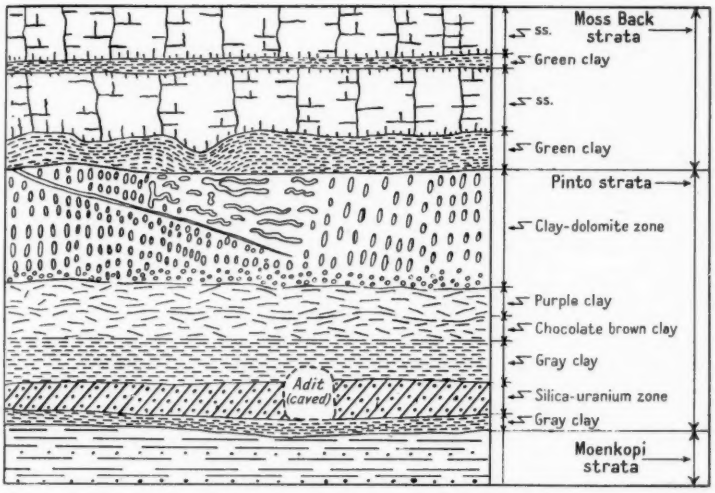

FIGURE 11.—DOLOMITE NODULES ABOVE CLAY ZONE

A sketch of the outcrop at Green Dragon No. 3, adit, San Rafael Swell. A portion of Figure 7 shown in greater detail.

The outcrop shown is about 43 feet high. The adit is approximately 7 feet high. The silicified layer ranges from massive quartz and chalcedony to chalcedony veinlets in altered clay.

Nodular dolomite is associated with haloes of clay alteration. Dolomite fills fractures and nodular forms show several directions of parallelism.

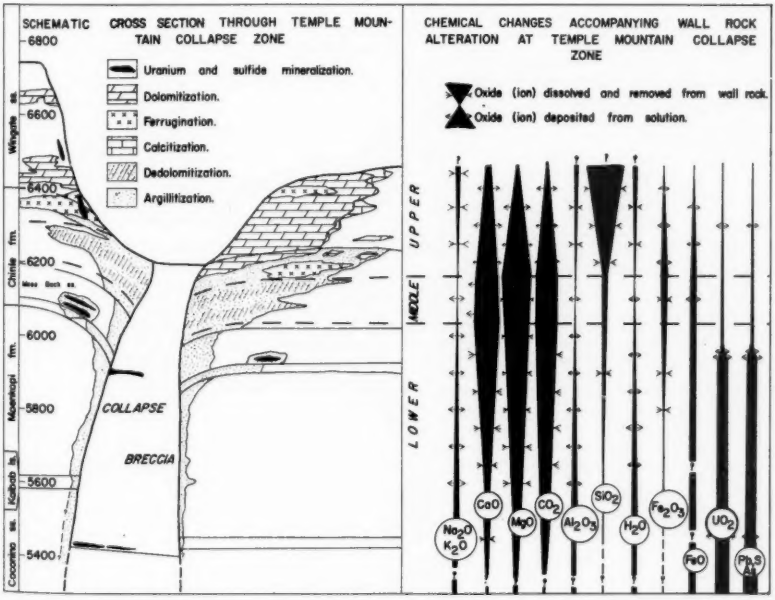

FIGURE 12.—SCHEMATIC DIAGRAM OF ALTERATION, TEMPLE MOUNTAIN COLLAPSE

The distribution of alteration phases is shown along the zone of collapse breccia. An indication of the accompanying chemical changes is shown at one side. The diagram is based on observations at Temple Mountian, Utah (after Kerr, Bodine, Kelley, Keys, 1957).

in a tuff with an accompanying diminution in radioactivity is the antithesis of the theory proposed for the plateaus whereby uranium would be leached from the tuff beds.

Alunite is found as a minor constituent of the uranium-bearing veins. Here it is presumably a late vein mineral possibly representing a more acidic phase.

Several occurrences of alunite have been mentioned in Plateau-type deposits. The mineral is associated with uranium in Capitol Reef and in the Happy Jack Mine.

While alunite is believed to be hydrothermal at Marysvale and is also recognized as a hydrothermal mineral at other localities, alunite of sedimentary origin has been found in lake deposits at Pidinga, South Australia (D. King, 1953), where the mechanism of formation is presumably lateritic representing an accumulation in old lagoons.

The occurrence of alunite in the Plateau uranium deposits may be associated with hydrothermal conditions, although the evidence of alunite alone is inconclusive.

Mineral Relationships

General review.—Mineral associations indicate both hypogene and supergene mineralization. Black, finely crystalline uraninite predominates in a number of deposits. Coffinite, a silicate of uranium, occurs frequently. Urano-organic ore is abundant. Uraninite deposits in a number of places contain common sulfides of copper such as chalcopyrite, bornite, chalcocite, and covellite. These may retain the cellular structure of wood and also exhibit exsolution effects of chalcopyrite after bornite. According to Takeuchi and Nambu (1956) bornite, which appears to be homogeneous at magnifications of 1000, segregates chalcopyrite in regular or irregular rods and blebs heated between 100° and 500° C. Therefore, the presence of this effect in Plateau uranium deposits may be a significant temperature indicator.

Numerous uranium minerals are supergene. Carnotite is probably the most widespread, but other minerals such as tyuyamunite, autunite, becquerelite, and schroeckingerite are common.

Among the many elements associated with Plateau uranium mineralization 11 may be selected as having a bearing on the study of origin: Se, Mo, As, Cu, K, V, Co, Cr, Pb, Zn, and S. The occurrence and distribution of the minerals containing these elements, while not indicative of any one unique interpretation, are compatible with original hydrothermal conditions.

Among key hydrothermal deposits in several major Precambrian areas the association of Co, Ni, Ag, and Au prevails in contrast to Cu, Pb, Zn, and V on the Plateau. The relative abundance of elements in the Precambrian group varies considerably. In the Belgian Congo deposit at Shinkolobwe (Derriks and Vaes, 1956), cattierite (CoS_2) and vaesite (NiS_2) are important associated minerals. Selenium-bearing sulfides also occur along with native gold. At the Great Bear Lake uranium locality of Canada, silver was mined extensively in the upper portions of certain veins, and Co and As have been consistently prominent as associated elements (Spence, 1931). At Joachimsthal, Czechoslovakia, Ni, Co, Ag, and As are prominent associated elements, although Cu, Pb, and Zn also occur (Everhard and Wright, 1953).

In Plateau deposits vanadium ranges from substantial amounts to traces. Copper is prominent in some areas, but absent from others. Selenium is a minor but widespread element. Arsenic and molybdenum are probably more widely disseminated than has been generally realized. Cobalt is present in a number of places but is erratic in distribution. Microscopic lead (galena) is present in many uraninite specimens. Zinc occurs in sphalerite in small amounts in several places.

Urano-organic ore.—The presence of urano-organic ore on the Plateau may indicate hydrothermal activity. The association of this ore with alteration and emplacement structures,

PLATE 1.—SILICIFICATION AND ALTERATION, TEMPLE MOUNTAIN, UTAH

FIGURE 1.—Chinle siltstone in Calyx cores, Temple Mountain, Utah.

FIGURE 2.—Silicified log in Moss Back strata, Temple Mountain, Utah. Silicified logs are non-ore-bearing.

PLATE 2.—DOLOMITE, SAN RAFAEL SWELL

FIGURE 1.—Massive dolomite, Temple Mountain, Utah. The dark rock exposures in the center of the view represent dolomite replacing quartzose sandstone.

FIGURE 2.—Dolomite nodules in siltstone. Replacements believed caused by hydrothermal solutions in Chinle strata on the west side of the San Rafael Swell, Utah, at Sulphur Springs Canyon.

FIGURE 1

FIGURE 2

SILICIFICATION AND ALTERATION, TEMPLE MOUNTAIN, UTAH

FIGURE 1

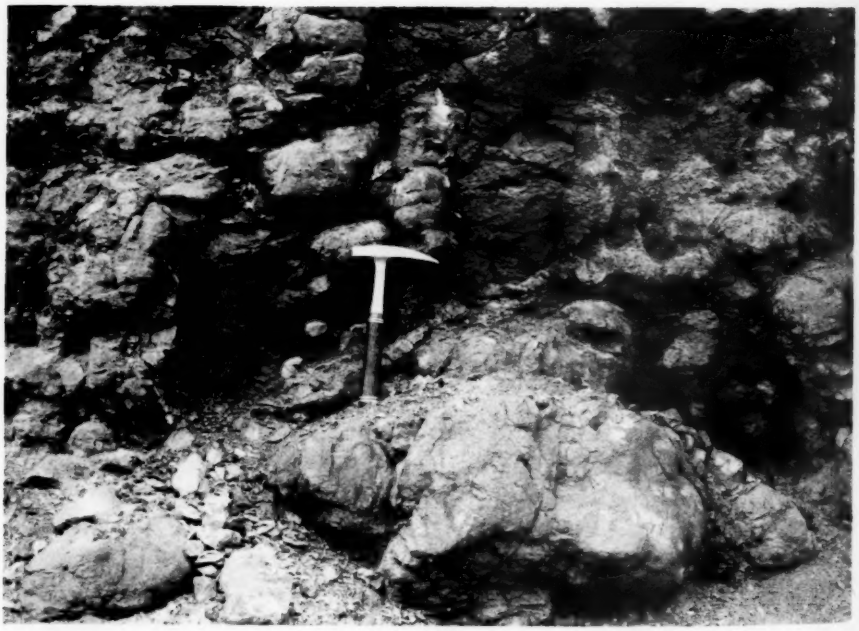

FIGURE 2

DOLOMITE, SAN RAFAEL SWELL

URANIUM-BEARING COLLAPSE NEAR A VOLCANIC VENT

as well as laboratory evidence, suggests this relation.

Both petroliferous and carbonaceous material are involved in the precipitation of uranium in Plateau strata. At Cameron, Arizona, the Happy Jack Mine, White Canyon, Utah, and the Delta Mine, Emery County, Utah, plant debris is replaced extensively by uraninite. At Temple Mountain, Utah, and Ambrosia Lake, New Mexico, petroliferous material appears to have been converted to urano-organic ore.

In the replacement of plant debris uranium-bearing solutions have permeated plant cells and caused partial or complete replacement by uraninite. Microscopic observation reveals well-preserved primary wood cell structures (Miller, 1955a). Accompanying hydrogen sulfide probably has been a contributing factor which caused the precipitation of uranium in surrounding porous sediments. At Cameron, Arizona, the smaller wood trash has contributed to ore formation whereas large silicified logs are barren. In carbonized logs at Temple Mountain, the uranium content is low or uranium is absent. On the other hand, the replacement of logs in the Delta Mine has produced fine uraninite specimens.

The occurrence of hard urano-organic material suggests the association of heated uranium-bearing solutions. Moore (1954) demonstrated experimentally that sub-bituminous coal removed 99.9 per cent of the uranium from a heated circulating uranyl sulphate solution. Kelley and Kerr (1958) show that natural semi-fluid, nonradioactive petroliferous tar also will extract uranium from a heated solution. A specimen of fluid tar from Temple Mountain, Utah, was refluxed in a uranyl sulfate solution at 104°C. and then heated for 96 hours at 280°C. The insoluble urano-organic residue contained 0.60 per cent equivalent U_3O_8.

Urano-organic ore precipitated by petroliferous material is believed to result when hydrothermal solutions carrying uranium ions penetrate the sands of an oil-bearing structure. The ore consists essentially of organic uranium compounds but contains microscopic uraninite, as well as some coffinite (Stieff *et al.*, 1956). The black and lustrous ore resembles hard asphalt. However, the volatile constituents are absent, the residue is insoluble in the usual organic solvents, its microhardness is greater than that of asphalt, and the material is uranium bearing.

Moulton's (1926) experiments show that the reduction of ferric to ferrous iron and the accompanying bleaching effect operate rapidly at atmospheric pressure by the action of crude oil vapor heated above 275°C. to the cracking temperature. Cracking temperatures vary, but Noller (1951, p. 59–60) reports that this process of thermal decomposition begins at 400°C. for the low-boiling-range gaseous and liquid hydrocarbons and continues to around 900°C. for the maximum yield of aromatic hydrocarbons. Natural clays act as catalysts for this process so that high-octane gasolines are obtained at 450–500°C. Thus, although the association of clay with hydrocarbon facilitates the cracking and bleaching processes, the experimental evidence indicates that at relatively low pressures heat is essential.

Collapse features and faulting at Temple Mountain occur on the southeast flank of the San Rafael Swell, a large oil-bearing anticlinal structure. Both collapse features and faults form breaks in the anticline. Bleaching of red-bed strata has accompanied both the collapse features and faults. Urano-organic ore is associated with the bleaching and extends outward from the bleached areas.

The urano-organic ores of Ambrosia Lake occur along the flanks of anticlinal structures which have been cut by small faults. A prominent ore associate is white kaolinite disseminated as spots through the ore. Indications of plant debris are negligible.

The natural cracking of organic material associated with a Tertiary sulfide vein is illustrated by a small deposit at La Bajada, New Mexico. Sulfide masses in veins exhibit fused fragments of organic matter which have been hardened and even converted to an amber-brown organic substance. Surrounding the fragments are clusters of spherical cavities (Fig. 13) caused by escape of volatiles. Vein deposition of sulfides, including pyrite, marcasite, and bravoite, is accompanied by hydrothermal argillic alteration of the wall rock. Some primary uranium mineralization is present.

A vein of urano-organic ore cuts Plateau

PLATE 3.—URANIUM-BEARING COLLAPSE NEAR A VOLCANIC VENT

A collapse in Chinle sandstone carries uranium mineralization. The circular area is slightly less than a mile in diameter and lies north of Shadow Mountain. The latter is the lava-encircled volcanic vent (black) in the lower left of the vertical aerial photograph. The area is about 10 miles northwest of Cameron, Arizona.

sediments at Placerville, Colorado. This material is similar to the organic accumulations at Temple Mountain, but the vein filling occurs along a fault. A combination of natural cracking, introduction of uranium, argillic alteration,

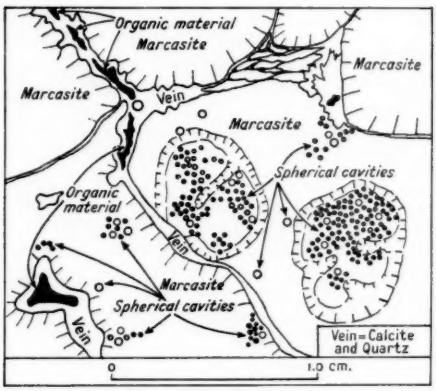

FIGURE 13.—SPHERICAL CAVITIES IN SULFIDE ORE
Volatilization of uranium-bearing organic material in veinlets is believed to have furnished cavity clusters. La Bajada, New Mexico.

and replacement of sandstone apparently was involved in the formation of the urano-organic ore.

Breger and Deul (1955) suggest that the organic constitutent of the ore at Temple Mountain represents an organic extract from sub-bituminous and lignitic coals. However, field observations do not appear to support this derivation from carbonaceous material. The amount of remnant lignitic material appears inadequate, and much ore occurs where plant debris is absent. Ore is found hundreds of feet both above and below the Moss Back member in which the plant detritus occurs. The deposition of organic material from mobile fluids is indicated by the occurrence of veinlets, lines of pellets along bedding planes, accumulations along fractures, and rounded masses which impregnate sandstone. The Cretaceous coal is widespread but stratigraphically much higher than this ore. The petroleum impregnation of porous sediments such as the Moss Back, which extends for miles beyond ore bodies, indicates the wide distribution of petroliferous material available for replacement.

The laboratory investigation of nodules by Breger and Deul (1955) shows that an increased specific gravity is accompanied by a rise in ash, total oxygen, nitrogen, and sulfur, but a decrease in carbon and hydrogen. Field observations (Kelley and Kerr, 1958) also indicate that the dense nodules contain more pyrite and uranium.

Microhardness determinations and differential thermal analyses indicate that as far as the physical and chemical nature of the organic material is concerned the urano-organic ore could have been derived from either a petroliferous or carbonaceous source (Kelley and Kerr, 1958).

Breger and Deul (1955) suggest that the organic matter might have been converted radiochemically to its present insoluble state. However, the mechanism for this theoretically significant conversion is unresolved. The energy of the atom is unquestioned, yet mineral changes which may be directly attributed to natural U^{238} essentially are minor. This postulated conversion would be slow since the tar which flows through cracks in the uranium ore at Temple Mountain is not radioactive. Furthermore, no destructive effects are observed in the vicinity of highly radioactive uraninite museum specimens kept in trays for many years. The well-known radioactive halos in biotite are local and usually microscopic in size.

Occasional sulfide nodules occur in the Temple Mountain ore, particularly in the collapse area (Fig. 14). They contain small urano-organic aggregates disseminated in pyrite, marcasite, native arsenic, and sphalerite. These nodules are 1–10 cm in diameter and replace quartz-bearing sediments. Some quartz has been reprecipitated with the sulfides and urano-organic constituent. These nodules therefore appear to have formed by precipitation at elevated temperatures.

Selenium.—Geobotanical prospecting on the Plateau has disclosed a widespread distribution of selenium. The element is a constituent of many plants found in the vicinity of uranium mineralization. Selenium may occur in hydrothermal deposits as shown by the selenium-bearing vaesite at Shinkolobwe in the Belgian Congo (Derriks and Vaes, 1956). It is associated with mercury as a hydrothermal replacement of limestone in the Lucky Boy deposit at Marysvale, Utah, where Plateau-type strata occur just west of the Plateau itself. The rare mineral tiemannite (HgSe) occurs in limestone (Kaibab ?); massive metallic specimens are found as float below the Lucky Boy Mine on the east slope of Deer Trail Mountain. The vein is no longer exposed.

Selenium has been reported as a constituent

of the uranium ores bordering the Henry Mountains (Butler, *et al.*, 1920) and in the ores of Silver Reef, Washington County, Utah. Selenium held in an unidentified compound is reported by Gruner and Knox (1957) at the Hardy Fee land, Converse County, Wyoming.

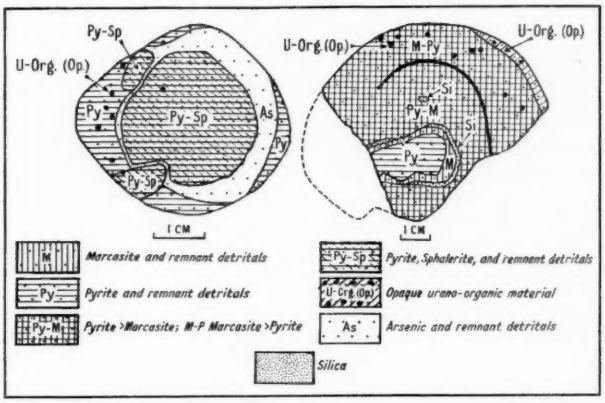

FIGURE 14.—SULFIDE NODULES IN URANO-ORGANIC ORE
Native arsenic, pyrite, marcasite, and sphalerite form nodular aggregates at Temple Mountain, Utah.

County, Colorado; and (3) Parco No. 23 Mine, Thompson's district, Grand County, Utah.

Selenium is widely distributed as a trace element in plants near selenium-bearing uranium deposits. Cannon (1953) reports that a selected sample of ore from the Westwater Canyon member of the Morrison formation at the Poison Canyon deposit, near Grants, New Mexico, contains 0.234 per cent selenium.

On the Plateau selenium is not only found close to uranium deposits in significant concentrations but may be dispersed for considerable distances by supergene action. Weeks, Coleman, and Thompson (1956) report native selenium, selenian pyrite, Fe $(S, Se)_2$ with as much as 5 per cent Se, and selenian marcasite, Fe $(S, Se)_2$ containing 1 per cent Se, on the Plateau. They also note clausthalite (PbSe) and eucairite (Cu Ag Se) on the Plateau. The primary source is not established, but the element is so widely distributed that it may be considered as part of the geochemical pattern.

Selenium occurs as a minor constituent of copper ores in the western United States. Needles of native selenium 2 cm long encrust quartzite and fire-baked sandstone at the United Verde Mine, Jerome, Arizona. These have been produced by the artificial fumarole conditions induced by a mine fire (Palache, 1934). The element is extracted commerically as a by-product of smelting from copper ore at Bingham Canyon, Utah.

Native selenium crystals have been identified by Thompson, Roach, and Braddock (1956) during their investigation of the mineralogy of uranium in sandstone-type deposits at (1) Black Hills, Fall River County, South Dakota; (2) Peanut Mine, Bull Canyon, Montrose

About 50 indicator plants, particularly genus *Astragalus*, contain Se, S, and Ca. They are reported (Cannon, 1957) to be associated with the selenium-bearing carnotite deposits on the Colorado Plateau, and their existence is attributed to the presence of near-by selenium mineralization. Beath (1943) has pointed out that the distribution of selenium-indicator plants is probably accordant with that of the uranium-vanadium deposits, and this relationship has been verified by mapping in the Yellow Cat district of Grand County, Utah, along the north flank of the Salt Valley anticline. The association of selenium with uranium over such a wide area is hardly a coincidence.

Molybdenum.—The widespread distribution of molybdenum is indicated by the frequency with which the often strikingly blue mineral ilsemannite ($Mo_3O_8 \cdot nH_2O$?) appears in mineral lists of the Plateau uranium deposits. "Jordisite," an amorphous (?) molybdenum sulfide which alters readily to ilsemannite, is less common. The rare mineral umohoite ($UO_2 \cdot MoO_4 \cdot 4H_2O$) was identified first at Marysvale, Utah, (Brophy and Kerr, (1953), where it is intimately associated with pitch-

blende in veins and is believed to be a primary low-temperature hydrothermal mineral. It is hydrated but loses water at a low temperature. It is interesting to note that molybdenite occurs with uranium at the La Veta Mine in this locality. Recently, a mineral similar to umohoite has been identified on the Plateau in a deep channel at Cameron, Arizona, where the relations also indicate a hydrothermal origin. Hinckley (1957, M. A. Thesis, Univ. Utah) reports that molybdenum is consistently more abundant in uranium-bearing areas in this locality. Umohoite has also been identified in the Gas Hills area in Fremont County, Wyoming, at the Lucky McMine (Coleman and Appleman, 1957). The authors attribute the origin at this occurrence to crystallization in the early stages of oxidation before the formation of the secondary uranium minerals with UO_2^{+2}.

The occurrence of molybdenum in hydrothermal veins which border the Plateau arouses interest in the occurrence of molybdenum minerals in the Plateau deposits, particularly since this element is distributed frequently in nature by hydrothermal activity. Ilsemannite has been reported by Botinelly, Weeks, and Johnson (1953) in the Monument No. 2 Mine; by Gruner and Gardiner (1952a) from Placerville, San Miguel County, Colorado, and White Canyon, Frey Canyon, and the Happy Jack Mine, San Juan County, Utah. The deposit at Placerville is in part a vein generally considered hydrothermal.

Molybdenum associated with tungsten is widely distributed in the intrusive areas of the western United States. In such occurrences, an origin connected with the aftermath of igneous activity has not been questioned. In view of this distribution, it is a reasonable possibility that the widespread dissemination of small amounts of molybdenum in the Plateau sediments may be due to upward migration from deeply buried originally igneous sources.

Arsenic.—Native arsenic is relatively abundant at Temple Mountain, Utah. It occurs disseminated and in nodules (Kerr and Lapham, 1954) in urano-organic ores associated with collapse features (Fig. 14). These nodules consist of a core of pyrite and sphalerite which replaces quartz, surrounded by a band of native arsenic, with shell of quartz, pyrite, and urano-organic ore. It is postulated that metallic arsenic was deposited at Temple Mountain from a heated liquid or gas. This hypothesis agrees with previously described occurrences. Evans (1903) attributed the vein occurrence at Montreal to fumarole action which started at certain points on the wall of the vein; successive layers of arsenic were added to produce an onionlike lump. The native arsenic at Washington Camp, Santa Cruz County, Arizona, is attributed by Warren (1903) to deposition by fumarole action in which the emanations may have escaped from underlying igneous rocks into a pocket in the limestone. Here, as a result of the local relief of pressure, layers of arsenic were deposited to form reniform masses.

Native arsenic has been obtained artificially (Doelter, 1918) as a sublimate from heated arsenopyrite precipitated by red-hot AgCl from a stream of CO_2 containing arsenic vapors. It is also produced by treating the arsenic sulfides, realgar and orpiment, with KOH, then heating with Na_2CO_3 to 300°C. The melting point of arsenic is 814°C. at 36 atmospheres (Mellor, 1929). Although a rock depth of 12,000 feet yields a static load which would lower the melting point or temperature of crystallization, it would hardly be lowered to 80°C., the temperature estimated to exist in ground-water conditions under this burden.

Realgar and some orpiment may be observed at Temple Mountain. Realgar is well known as a sublimate on the cooler surface layers at Vesuvius and Etna. According to Seebach (1926) it is deposited by fractional distillation at temperatures of 210–270°C. around fumaroles at Papandayan, Java. Realgar also occurs with siliceous sinter at Yellowstone Park. This sulfide has formed from burning coal in mine fires and on mine dumps at several localities, including the Plaunenschen Grunde area around Dresden, East Germany, at Silesia, Poland, and in the French Central Plateau. Seebach reports (1926) that prismatic crystals of realgar are obtained on the glass walls by heating realgar or artificial AsS to 150°C in a closed tube with an excess of sulfur in a solution of sodium bicarbonate.

The occurrence of domeykite (Cu_3As) with tennantite [$(Cu, Fe)_{12}As_4S_{12}$] on the plateau along with uranium mineralization is consistent since this mineral usually is associated with Cu, Pb, and Zn in low- to moderate-temperature hydrothermal veins. Incomplete X-ray-diffraction patterns of specimens from Temple Mountain suggest arsenopyrite, but recognizable crystals were not observed (Kelley and Kerr, 1958). At least 10 arsenic-bearing minerals (Table 2) occur on the Plateau with uranium mineralization (Weeks, Coleman, and Thompson, 1956; Gruner and Gardiner, 1952a; Thompson, Ingram and Gross, 1956).

Traces of arsenic occur with the organic constituents of sedimentary rocks (Onishi and Sandell, 1955), and traces have also been reported from petroliferous materials (Erickson, Myers, and Horr, 1954). However, the arsenic content at Temple Mountain represents a wide

TABLE 2.—ARSENIC MINERALS OF THE PLATEAU

native arsenic	As
domeykite	Cu_3As
realgar	AsS
orpiment	As_2S_3
gersdorffite	NiAsS
tennantite	$(Cu, Fe)_{12}As_4S_{13}$
metazeunerite	$Cu(UO_2)_2(AsO_4)_2 \cdot 8H_2O$
zeunerite	$Cu(UO_2)_2(AsO_4)_2 \cdot 10\text{–}12\ H_2O$
abernathyite	$K_2(UO_2)_2(AsO_4)_2 \cdot 8H_2O$
novacekite	$Mg(UO_2)_2(AsO_4)_2 \cdot 8\text{–}10H_2O$

vertical range and is attributed to deep-seated sources rather than to concentration from near-by organic sources.

Copper.—The copper-mineral associations of the Plateau, the distribution of copper minerals along some channels and not along others, and the criteria of replacement indicate active copper-bearing solutions. The similarity between copper minerals observed in Plateau deposits and minerals of hydrothermal copper veins suggests a similar origin. Table 3 lists 28 copper-bearing minerals for the Plateau (Gruner and Knox, 1957; Keys, 1954; Weeks, Coleman, and Thompson, 1956). Trites, Chew, and Lovering (1955) have published an extensive mineral list from the Happy Jack Mine. Everhard (1950) has examined the copper-uranium deposits of Silver Reef. Brooke, Shirley, and Swanson (1951) in an investigation of the Salt Wash in the trachyte district east of the Henry Mountains, Utah, report sulfides believed to be pyrite and bornite in an adit on Farmer's Knob. At Capitol Reef, Utah, Smith, Hinrichs, and Luedke (1952) report that zippeite and metatorbernite are associated with copper mineralization.

Chalcopyrite is prepared by the fusion of pyrite and copper sulfide and by gently heating cupric and ferric oxides in an atmosphere of H_2S. Schwartz (1927) produced chalcopyrite at 550°C. from an intergrowth of chalcopyrite and bornite. Covellite has been synthesized by Posnjak, Allen, and Merwin (1915) from cuprous sulfide or metallic Cu in an atmosphere of H_2S at 358°C. Amorphous cupric sulfide is also precipitated with H_2S from a 2.5 per cent solution of copper sulfate containing 1 per cent sulfuric acid. After heating in H_2S to 150°C. the material appears to crystallize. The association of pyrite and these copper sulfides indicates the introduction of minerals under hydrothermal conditions perhaps with accompanying H_2S.

Copper is an important accessory element in

TABLE 3.—COPPER MINERALS OF THE PLATEAU

chalcopyrite	$CuFeS_2$
bornite	Cu_5FeS_4
tetrahedrite	$(Cu, Fe)_{12}Sb_4S_{13}$
tennantite	$(Cu, Fe)_{12}As_4S_{13}$
domeykite	Cu_3As
covellite	CuS
chalcocite	Cu_2S
digenite	$Cu_{7-x}S$
euchirite	CuAgSe
malachite	$CuCO_3 \cdot Cu(OH)_2$
azurite	$2CuCO_3 \cdot Cu(OH)_2$
chalcanthite	$CuO \cdot SO_3 \cdot 5H_2O$
antlerite	$3CuO \cdot SO_3 \cdot 2H_2O$
brochantite	$4CuO \cdot SO_3 \cdot 3H_2O$
boothite	$CuSO_4 \cdot 7H_2O$
chalcoalumite	$CuAl_4(SO_4)(OH)_2 \cdot 3H_2O$
cyanotrichite	$Cu_4 \cdot Al_2(SO_4)(OH)_{12} \cdot 2H_2O$
johannite	$Cu(UO_2)_2(SO_4)_2(OH)_2 \cdot 6H_2O$
torbernite	$Cu(UO_2)_2(PO_4)_2 \cdot 8\text{–}12H_2O$
metatorbernite	$Cu(UO_2)_2(PO_4)_2 \cdot 6\text{–}8H_2O$
zeunerite	$Cu(UO_2)_2(AsO_4)_2 \cdot 10\text{–}12H_2O$
metazeunerite	$Cu(UO_2)_2(AsO_4)_2 \cdot 8H_2O$
conichalcite	$CaCu(AsO_4)(OH)$
chrysocolla	$CuSiO_3 \cdot 2H_2O$
cuprosklodowskite	$Cu(UO_2)_2(SiO_3)_2(OH)_2 \cdot 5H_2O$
calciovolborthite	$(Cu, Ca)_2(VO_4)(OH)$
volborthite	$Cu(VO_4)_2 \cdot 3H_2O$
vesignieite	$Cu_2Ba(VO_4)_2(OH)_2$

at least five uranium deposits (Botinelly and Weeks, 1957) for about 65 miles along the Colorado River. Elsewhere, it is sporadic and less significant. The deposits high in copper characteristically are nonvanadiferous and unoxidized. Miller (1955b) has reported the replacement of wood by chalcopyrite (Fig. 15) which in turn is replaced by pitchblende at the Happy Jack Mine. The relation is accepted by some as indicative of hypogene mineralization. Dodd (1950) reaches this conclusion.

Potassium.—The distribution of the secondary potassium-bearing mineral, carnotite, in the Colorado Plateau suggests a regional relation to the underlying saline strata of the Pennsylvanian Paradox Basin. Since a primary potassium-bearing uranium mineral has not been reported, the source of this element is of considerable interest. It is thought that this secondary mineralization could be related to circulating low-temperature ground waters enriched with potassium derived from nearby-salt plugs and saline strata. Harrison (1927) reports drilled wells which have penetrated

several hundred feet of salt and anhydrite gypsum cap rock in the saline anticlines and plugs along the Utah-Colorado boundary in Grand and Montrose counties. This is in the vicinity of significant carnotite mineralization.

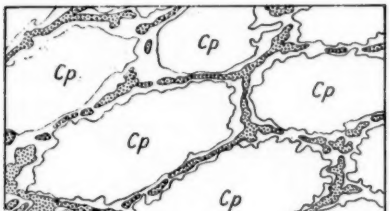

FIGURE 15.—URANIUM ORE REPLACING A LOG
A polished surface of ore from the Happy Jack Mine (courtesy of L. J. Miller) shows chalcopyrite filling cells and uraninite replacing cell walls.

At Moab and Salt Creek these saline accumulations form prominent topographic features, whereas the large area of deeper-seated salt lies to the south and west. This region has been drilled extensively by several oil companies.

That these salt plugs may contain significant amounts of potassium is shown by Lindberg (1946) who described two cores drilled near Thompsons, Utah. They consist of a relatively simple mineral assemblage throughout a distance of 547 and 813 feet. Although halite is the most prominent mineral, carnallite ($KMgCl_3 \cdot 6H_2O$) and sylvite (KCl) are abundant. Minor amounts of kieserite, anhydrite, and clay also are associated. The total carnallite and sylvite content of specimens from these two cores ranges between 7 and 87 per cent. Chemical analyses show that the K_2O content derived from these two readily soluble minerals ranges between 2.38 and 26.60 per cent. This gives an overall average of 9.77 per cent K_2O throughout the length of the two cores.

It seems reasonable to conclude that the contribution of potassium to circulating ground water in this region has been appreciable and could be a factor in the processes which led to the development of the widely disseminated secondary carnotite. The fact that this material is localized in the upper supergene zones rather than in the deeper primary zones lends credence to the theory that these deposits are related to circulating potassium-bearing ground waters.

Vanadium.—A broad arcuate belt containing uranium deposits high in vanadium largely in the Salt Wash member of the Morrison formation extends for about 100 miles across eastern Utah and western Colorado. Weeks, Coleman, and Thompson (1956) report 23 vanadium minerals from Colorado Plateau uranium localities. The deposits are largely highly oxidized, but in some occurrences (roscoelite and montroseite) oxidation has been partial or negligible. The application of the hydrothermal theory to the original constituents of the Salt Wash has encountered difficulty because of the absence of a primary mineral subject to alteration to carnotite. Rasor (1952) in a description of massive uraninite from the Grey Dawn Mine, San Juan County, Utah, pointed to the presence of an associated unidentified black mineral high in vanadium. Such a mineral suggests a primary source for the vanadium.

TABLE 4.—VANADIUM-URANIUM-BEARING MINERALS OF THE PLATEAU

uvanite	$U_2V_6O_{21} \cdot 15H_2O$(?)
carnotite	$K_2(UO_2)_2 \cdot (VO_4)_2 \cdot 1\text{–}3H_2O$
metatyuyamunite	$Ca(UO_2)_2(VO_4)_2 \cdot 3\text{–}5H_2O$
tyuyamunite	$Ca(UO_2)_2 \cdot (VO_4)_2 \cdot 5\text{–}8\frac{1}{2}H_2O$
rauvite	$CaO \cdot 2UO_3 \cdot 2V_2O_5 \cdot 16H_2O$(?)

On the basis of essential mineral content, Finch (1953) has described three types of Shinarump deposits as vanadium-uranium, copper-uranium, and uranium. The predominant minerals are (1) carnotite and tyuyamunite; (2) uraninite, chalcocite, coveliite, and secondary copper-uranium minerals; and (3) uraninite with few secondary minerals. Finch has pointed to the gradual diminution of the vanadium content of ores westward. Data on the uranium and vanadium minerals of the Plateau are being assembled for early publication by A. D. Weeks and M. E. Thompson of the U. S. Geological Survey. Common vanadium-uranium minerals are given in Table 4. Eighteen vanadium minerals are found in which uranium is not a constituent element. These consist of oxides, vanadates, and silicates and are all potential associates of uranium ores.

Carnotite and tyuyamunite are found particularly in the Colorado-Utah Four Corners area, in Wyoming, and in South Dakota. Rauvite is less widespread and uvanite has been found only at Temple Mountain. At the Mi

Vida Mine near Moab, Utah, vanadium appears to be more abundant around the margins of the uranium ore body. A zonal arrangement also is found at Temple Mountain.

In the Morrison deposits vanadium has been subject to extensive supergene deposition, yet the occurrence of montroseite, a hydrothermal vanadium mineral, indicates an introductory stage in contrast to normal supergene distribution. In places the possibility of primary hydrothermal introduction exists.

Cobalt.—Uranium deposits contain small amounts of cobalt (Table 5) at several places

TABLE 5.—COBALT-BEARING MINERALS OF THE PLATEAU

cobaltian pyrite	$(Fe, Co)S_2$
bieberite	$CoSO_4 \cdot 7H_2O$
unnamed	$CoSO_4 \cdot 6H_2O$
unnamed	$(Co, Fe)SO_4 \cdot 4H_2O$
unnamed	$(Co, Fe)SO_4 \cdot 6H_2O$
erythrite	$(Co, Ni)_3(AsO_4)_2 \cdot 8H_2O$
cobaltian pickeringerite	$MgAl_2(SO_4)_4 \cdot 22H_2O$
cobaltocalcite	$CoCO_3$
lithiophorite	Co, Mn wad

(Keys, 1954; Weeks, Coleman, and Thompson, 1956; Gruner and Gardiner, 1952b; Gruner and Gardiner, 1952a)

on the Plateau. The element has been reported in a pink oxidation product at the Lucky Strike Mine on the San Rafael Swell and in bieberite and cobaltian wad at Cameron, Arizona. It appears to be much more local in distribution than molybdenum.

Colbatian pyrite $(Fe,Co)S_2$ with as much as 1 per cent Co is reported by Weeks, Coleman, and Thompson (1956) as a Colorado Plateau accessory mineral. They note two generations of pyrite in the primary uraninite ores at the Delta Mine, Emery County, Utah (p. 25) "...with the pyrite in the uraninite showing rims of unidentified nickel and cobalt minerals—further evidence of nickel-cobalt enrichment accompanying the uranium mineralization."

Gersdorffite (NiAsS) is reported at the Happy Jack Mine (Gruner and Gardiner, 1952a). This uncommon mineral ordinarily occurs typically in veins with other nickel minerals such as niccolite, ullmannite, and chloanthite.

Chromium.—The occurrence of bright-green chrome mica clay associated with uranium mineralization at Temple Mountain, Utah, and Placerville, Colorado, is described by Kerr and Hamilton (1958). McConnell (1953) reports volkonskoite, chromium montmorillonite, at Thompsons, Utah, where it occurs in high-angle fissures and is associated with uranium mineralization. At Temple Mountain this clay contains between 0.4 and 3.6 per cent Cr_2O_3 and consists of mixed $2M_1$, 1M mica polymorphs. It is thought to represent a temperature of formation somewhere between 200°C. and 350°C. The chrome mica clay at Placerville also consists of mixed $2M_1$, 1M polymorphs, but in addition contains interlayered montmorillonite. In both localities the chrome mica clays appear to represent an impregnation due to solutions associated with the introduction of uranium.

Well-crystallized chrome micas, also known as fuchsite and mariposite, are also associated with the Canadian Shield and California Mother Lode gold deposits. Whitmore, Berry, and Hawley (1946) conclude that the majority of these chrome-mica occurrences associated with gold are conspicuous along faults and strongly suggest a derivation from hydrothermal solutions rising from magmas. The associated minerals indicate that the chrome micas formed under conditions of moderate to fairly high temperatures and pressures.

The poorly crystallized, mixed polymorph mica clays associated with Plateau uranium mineralization seem to reflect lower temperatures of formation than the crystalline chromium micas associated with the hydrothermal gold deposits.

Lead.—Galena is found in small isolated veinlets in strata at the base of the Temple Mountain collapse (Kerr, Bodine, Kelley, and Keys, 1957). It occurs in isolated granular masses at the Orphan Mine, Grand Canyon, Arizona. Isolated grains an eighth of an inch in diameter have been observed in massive uraninite from the Mi Vida Mine. Such occurrences probably represent direct precipitation from solution.

In addition, galena is present in minute particles in many of the uraninite specimens collected from different localities on the Plateau. The distinction between the extent to which such galena represents radiogenic origin and the extent to which it represents direct precipitation is often difficult.

Sulfur.—Sulfur in pyrite and copper sulfides may have been derived from associated organic material and in part introduced into sediments.

North of Sulphur Creek in the Captiol Reef area, Wayne County, Utah, Smith, Hindrichs, and Luedke (1952) report pyrite nodules as large as 2 by 3 by 6 inches in a zone 10 feet thick. The sandstone matrix is white and contains small grains of green and blue copper minerals.

Allen, Crenshaw, Johnson, and Larsen (1912)

synthesized mixtures of pyrite and marcasite by the action of hydrogen sulfide on ferric sulfate at temperatures between 100°C. and 300°C. At 200°C. as the acid (H_2SO_4) concentration increases, the percentage of marcasite increases, whereas at the same pH the percentage of pyrite with respect to marcasite increases as the temperature rises from 100°C. to 300°C.

TABLE 6.—VERTICAL SPREAD OF HOST ROCKS OF PLATEAU-TYPE URANIUM OCCURRENCES*

Age	Area	State	Authority
Pliocene	Split Rock	Wyoming	Love (1953)
Miocene	Miller Hill	Wyoming	Vine and Prichard (1956)
Oligocene	Harding and Pennington Counties	South Dakota	Gill (1953); Moore and Levish (1955)
Eocene	Gas Hills	Wyoming	Love (1953)
	Lost Creek	Wyoming	Page (1950); Wyant (1952)
	Pumpkin Buttes	Wyoming	Love (1952)
Cretaceous	Gas Hills	Wyoming	Love (1953)
	Edgemont	South Dakota	Gott, Page, and Jones (1953)
	Alladin	South Dakota	Gray and Tennissen (1953)
	Gallup	New Mexico	Rapaport *et al.* map (1952)
	Orderville	Utah	Beroni *et al.* (1953)
Jurassic			
Salt Wash	Plateau	Colorado-Utah	Craig *et al.* (1951)
Westwater	Plateau	New Mexico	Page, Stocking, and Smith (1956)
Todilto	Plateau	New Mexico	
Triassic			
Shinarump	Plateau	Utah-Arizona	De Vergie and Carlson (1953)
(Moss Back)	Plateau		
Mississippian	Marysvale	Utah	Bethke and Kerr (1954)
Cambrian	Lawrence County	South Dakota	Vickers (1953)

* Partial list subject to revision.

Stratigraphic Distribution

The uranium-bearing rock units on the Colorado Plateau include 32 formations which range in age from Pennsylvanian to Tertiary. Of these, 15 have produced more than 1000 tons of ore (Isachsen, Mitcham, and Wood, 1955). The widespread vertical occurrence is evidence for uranium emplacement from external sources, although ore in specific areas is restricted to a few particularly favorable formations.

A complete stratigraphic review of Plateau uranium occurrences is beyond the scope of this paper, but a few are selected to illustrate more specifically the range in wall-rock ages (Table 6).

Pliocene strata of the Split Rock area along the Fremont-Natrona County boundary in Wyoming yield small amounts of radioactivity (Love, 1952). The Browns Park formation, probably Miocene, in Carbon County, Wyoming, at Miller Hill (Vine and Prichard, 1956) contains uranophane (?) in the upper part of limestone strata and in calcareous sandstone beds. The uranium is attributed to leaching from younger tuffaceous sandstone followed by precipitation along with free silica in lower strata.

Strata of Oligocene age in Harding and Pennington counties, South Dakota, contain uranium. Gill (1953) reports carnotite in tuffaceous sandstone of the Chadron formation of the White River group, and Moore and Levish (1955) report uranocircite and metatyuyamunite (?) in the limestone facies of this unit.

Love (1954) reports uranospinite, $Ca(UO_2)_2(AsO_4)_2$ $8\text{–}10H_2O$, from the Gas Hills area along the Fremont-Natrona county boundary Wyoming. The mineral occurs in argillic and conglomeratic sandstone and in carbonaceous shale. The uranium-bearing strata lie in the Wind River formation (early Eocene) and the Thermopolis shale (early Cretaceous).

Page (1950) and Wyant (1952) have reported on the schroeckingerite deposits in the Lost Creek area, Sweetwater County, Wyoming. Clay strata containing uranium are said to be Lower Eocene or younger and are interbedded with arkosic sands and grits which contain disseminated grains of the uranium mineral. Several sources are tentatively considered: (1) uranian lignite beds possibly 300–500 feet below; (2) primary vein material occurring in depth along the Cyclone River fault zone; (3) sediments derived from surrounding granites, gneisses, and schists; or (4) possibly through evaporation of lake water along an old shore line.

Love (1952) has reported the occurrence of uranium in several types of sandstone and at several stratigraphic horizons in the Wasatch formation (Lower Eocene) of the Pumpkin Buttes area, Campbell County, Wyoming. He postulates an origin dependent on leaching uranium from tuffs in the White River formation above and precipitation in the Wasatch strata below.

Gott, Page, and Jones (1953) report a need for further stratigraphic study to clarify the boundaries in the formations of the Edgemont area of the Southern Black Hills, South Dakota, in which uranium is found. The Dakota, Fuson, and Lakota formations from the area, however, are generally recognized as Cretaceous. Botinelly, Weeks, and Johnson (1953) report that the ores of the Edgemont area contain tyuyamunite, rauvite, hewettite, corvusite, barite, and native selenium, a mineral association similar to that found in the oxidized vanidiferous ores of the Colorado Plateau.

On the northern side of the Black Hills uplift, near Alladin, Crook County, Wyoming, Gray and Tennissen (1953) report 42 radioactive anomalies connected with uranium mineralization in varying amounts. Carnotite, torbernite (?), and autunite are the minerals mentioned. The mineralization is found at three stratigraphic elevations in the Fall River formation, in the Fuson shale, and in the Lakota sandstone—all Cretaceous.

At the northwestern end of the Zuni uplift in New Mexico, on the outskirts of Gallup, uranium has been found in black organic shale. The uranian strata have been mapped as Dakota (Cretaceous) by Rapaport, Hadfield, and Olson (1952).

Northwest of Orderville in Kane County, Utah, uranium deposits have been described by Beroni, McKeown, Stugard, and Gott (1953). The occurrence is in strata previously mapped by Gregory (1950) as Dakota (Cretaceous).

Craig *et al.* (1951) state that the deposits of the Morrison formation (Jurassic) are largely confined to the Salt Wash member and lie essentially within the sandstone and mudstone facies of the Salt Wash. In local areas, however, the uranium-bearing zones may change by steps to a considerable degree. Although the beds are ordinarily considered as carnotite bearing, other minerals are recognized, and even so-called "core" areas of pitchblende are found in a number of places.

The Entrada sandstone (Jurassic), a light-colored massive cliff-forming formation exposed in western Colorado and eastern Utah, is listed by Craig *et al.* as vanadium-bearing.

Certain portions of the Carmel formation (Jurassic) in eastern Utah and western Colorado are known to be uranium bearing.

The Chinle formation is recognized as one of the most widely distributed of the Plateau units. As indicated by Colbert (Personal communication), it has long furnished collecting areas for dinosaurs. The Shinarump conglomerates and sandstone beds at the unconformity with the Moenkopi and occasionally some distance above yield some of our most important uranium deposits. Steward and Williams (1954) point out that in White Canyon the Shinarump conglomerate is a well-developed coarse-grained sandstone that probably correlated with the Shinarump of the type section in northwestern Arizona. In the area between the Green and Colorado rivers and terminating north of Moab, the unit called Shinarump sandstone and conglomerate has been tentatively named the Moss Back sandstone and represents a horizon above the true Shinarump. In the southern part of the San Rafael Swell the Shinarump (Moss Back) correlates with the uppermost member of a so-called "D" unit in the Chinle formation. The Moss Back sandstone is about 200 feet higher stratigraphically than the Shinarump in the White Canyon area. A lack of correlation between ore-bearing conglomerates at the base of the Chinle in the area of the Colorado and Green rivers has also been recognized by De Vergie and Carlson (1953).

The "C" group uranium deposit in Shinarump conglomerate along the Colorado River below Moab, Utah, cannot, it is claimed, be correlated directly with the Shinarump in the White Canyon area to the south or with that of the Dead Horse Point area to the north.

Major uranium deposits have not been

reported in Paleozoic strata of the Plateau area. Scattered occurrences, however, indicate the possibility of such occurrences.

At Red Bluff, Arizona, uranium is found along fractures in Precambrian quartzite of the Dripping Springs formation as reported by Granger and Raup (1956). Kaiser (1951) had reported disseminated ore in the prospect and at the same time favored hydrothermal origin.

In the northeastern part of the Wind River Basin, Wyoming, Tourtelot (1952) reported 31 radioactive anomalies. Of these, 22 are in Precambrian granite, 5 are in middle and upper Eocene rocks, 2 are in alluvium, 1 is in Precambrian schist and gneiss, and 1 is in Flathead sandstone.

Vickers (1953) has reported that autunite occurs as fracture coatings and disseminations in siltstone of the Deadwood formation (Cambrian) in Lawrence County, South Dakota. The mineralization is concentrated mainly in the lower 2 feet of the siltstone at the contact with intrusive rhyolite porphyry. The gangue minerals are fluorite and limonite.

The range in occurrence from Pliocene to Precambrian would present a difficulty under the ash-leach theory. Adequate pyroclastic distribution at sufficient age levels to supply sources does not appear to exist. Likewise, the requirements of syngenetic origin fail to satisfy this extended stratigraphic range.

Structural Factors in Emplacement

Notwithstanding the use of the term "Plateau," uranium deposits are seldom found in geometrically horizontal strata. Many deposits occur in strata encircling eroded anticlinal structures or along the escarpments of monoclinal uplifts. Few, if any, sizeable deposits are found where structural disturbance is absent.

Long ago Gilbert (1877) pointed out that buried laccoliths or other intrusive forms may be related to structural features observed at the surface. Later Butler *et al.* (1920) emphasized the intimate genetic relation between the structural features of the region and the ore deposits.

A structural study of the Zuni uplift led Bucher and Gilkey (1953) and Gilkey (1953) to conclude that the monoclinal folds observed at the surface are probably related to deformation in crystalline rocks far below. Blocks of the earth's crust, bounded by major fractures and extending upward, stretch the sediments draped over them.

In Bucher's opinion, upward bulging of the strata accompanies fracturing. Further, the differential movement caused by continued bulging is in a vertical direction, although the fractures develop as a consequence of the horizontal stresses set up in the surface.

Many local uranium districts are recognized

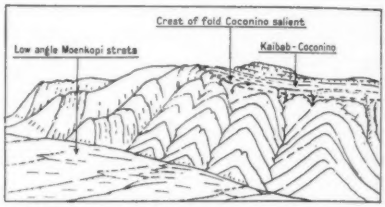

FIGURE 16.—MONOCLINAL FOLD IN COCONINO SALIENT

Southwest of Cameron, Arizona. Nearly flat Kaibab strata on the ridge and flat Moenkopi below are separated by a monoclinal flexure.

in terms of the associated regional structures. The structural features of the Colorado Plateau have been summarized by Shoemaker and Luedke (1952) from the works of Baker (1946), Babenroth and Strahler (1945), Darton (1925), and others.

In New Mexico one finds the Zuni and Lucero uplifts, the eastern margin of the Defiance uplift, the Hogback Mountain monocline, the Nacimiento uplift, and the San Juan Basin. In Arizona, encircling Black Mesa basin, are the Defiance uplift, the Carrizo Mountains, the Comb Ridge monocline, Navajo Mountain, Echo Cliffs monocline, the East Kaibab monocline (Fig. 16), and the Kaibab uplift.

In Utah, the East Kaibab and Echo Cliffs monoclines lie west and south of the Kaiparowitz Basin; Navajo Mountain, the Circle Cliffs uplift, the Waterpocket Fold, and the Henry Mountains lie to the east and northeast. In the southeastern corner of Utah are the Monument uplift, the Comb Ridge monocline, and the Abajo Mountains. Extending northward from the Henry Mountains is the San Rafael Swell.

One of the most interesting structural areas surrounds the La Sal Mountains in eastern Utah and western Colorado. A series of northwesterly trending structures includes the Lisbon Valley and Dolores anticlines, the Gypsum Valley anticline, the Moab Valley anticline, the Castle Valley and Paradox Valley anticlines, the Salt Valley-Fisher Valley-Sinbad Valley anticline, and the Uncompahgre uplift. Along the eastern margin of the Plateau in Colorado are the La Plata, Rico, and San

Miguel Mountains, the Gunnison uplift, the West Elk Mountains, and the Grand Hog Back monocline.

In Wyoming, uranium is found in the Wind River Basin and the Powder River Basin. Along the flanks of the Black Hills uplift the uranium-producing area extends into South Dakota.

In almost every uplift, anticlinal, or monoclinal structure in the region outlined uranium mineralization has been found.

Evidence of fracture influence on uranium deposition may be noted in the Zuni uplift. At Gallup, New Mexico, black uranium-bearing shales appear to have received uranium mineralization through fractures in underlying sandstone. The shale lies in the Dakota formation in finely fractured strata about 15 feet thick and exposed by an open cut for a distance of about 1000 feet along the strike. The strata are inclined at about 45°, and a coarse fracture pattern is seen in the sandstone beneath. The basal portion of the shale just above the sandstone is highly radioactive, while the shale higher in the section becomes less and less radioactive. Further, radioactivity is concentrated not only along the contact but close to fractures that cut the underlying sandstone. It is easily conceivable that the distribution of radioactivity observed could be due to impregnation in the shale caused by uranium-bearing solutions migrating upward through fractures in the sandstone.

Both Bucher and Gilkey (1953) and Gilkey (1953) report a parallelism between fracture trends in the Todilto limestone and the strike of fluorite veins in crystalline rocks which occupy the center of the Zuni uplift. This places emphasis on the fluorite association with the uranium ores of the Todilto limestone, a feature now well recognized.

A feature of consuming interest to all who have observed it is the tongue of light-gray broken strata which extends down the west slope of Temple Mountain in sharp color contrast to the red unbroken strata of the Moenkopi and Chinle formations. The area of deformation has been referred to as the "Flop-over" by uranium miners. It was described as a zone of hydrothermal alteration by Hess (1922). Later Gruner (1951) reported, "The structure is unmistakably a 'double' landslide, one to eastward the other westward, of white Wingate sandstone." Later however, Gruner, Gardiner, and Smith (1953) revised Gruner's earlier opinion and referred to the feature as a "collapse" structure.

During the summer of 1953 several additional collapse areas were observed in the vicinity of Temple Mountain. Four occur west and southwest and two northeast of the main Temple Mountain area, all within an area of 8–10 miles. Faulting traceable for a considerable distance is associated with the areas to the north and southwest. At least 10 collapse areas are now known in the southern part of the San Rafael Swell.

In the near-by occurrences strata traceable without interruption on either side are downdropped within the broken areas. The areas on the southwest may each cover a few acres, while one large area on the north covers about half a section. In each a sharp color contrast occurs between the almost white bleached rock of the collapse and the bordering red Moenkopi. Sands are impregnated with hydrocarbons in each area. Conglomerates indicate that at least some of the broken strata involved were derived from the Moss Back. In each place the "collapse" occurs below the level of the Moss Back and penetrates the Moenkopi.

The mechanism of the collapse is difficult to explain without the removal of a considerable underlying rock mass. The most likely substance for such removal would be the salines from the Paradox formation known to be some distance stratigraphically below. An unusual combination of circumstances, however, would be required to remove the salines from only some areas. Faulting extending downward to salt strata elevated above the water table might provide channels for percolating meteoric waters which would act intermittently, ultimately resulting in collapse. This explanation, however, would not account for the bleaching and the distribution of hydrocarbons in the collapsed area. Neither would it explain the uranium deposit in the "collapse" at Temple Mountain itself.

The minerals of the deposit in the "collapse" at Temple Mountain include polymerized hydrocarbon with a high uranium content and copper sulfides. Also, a number of pyrite nodules have been found. One, about an inch in diameter, was encased within an outer layer of native arsenic. Some sphalerite is mixed with the pyrite in the core of the nodule. The conditions observed are in part at least consistent with a hydrothermal origin as first suggested by Hess (1922).

Faults of considerable displacement have been noted in various parts of the Plateau area. Even uplifts such as the San Rafael Swell are cut by faults (Fig. 17). Few portions of the

Plateau of any considerable magnitude are free from faulting. Some faults are connected with mineralization. A notable example is the faulting near Placerville, Colorado (Kerr, Rasor, viewed from reconaissance plane flights. Intermediate thin-bedded strata may be fractured according to a finely textured pattern.

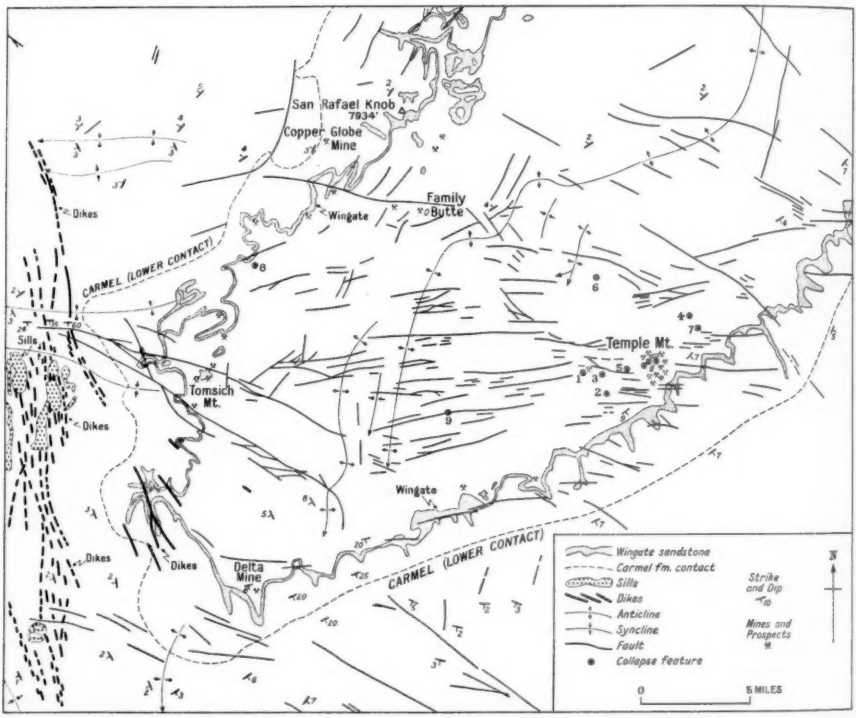

FIGURE 17.—STRUCTURAL PATTERN, SOUTHERN SAN RAFAEL SWELL, UTAH
The southern portion of the Swell is transected by a superimposed east-trending fault pattern. To the west swarms of north-trending dikes cut Plateau strata. Prominent northwest faults lie between.

and Hamilton, 1951), where a vein containing radioactive hydrocarbon, specks of uraninite, and sulfide minerals such as chalcopyrite, tetrahedrite, sphalerite, and pyrite has formed along a fault plane which cuts the sedimentary strata. This has been accepted as hydrothermal in origin by Gruner (1951).

In a few places such as Red Bluff in Arizona and near Edgemont in the southern Black Hills of South Dakota uranium minerals have formed along fractures in sedimentary strata. At Silver Reef in Utah fractures have been formed on a broad scale in the ore-bearing zone and in the Shinarump formation below.

Coarse fracture patterns in the more massive beds of Plateau sediments may be observed in uranium-bearing areas, particularly when

On the basis of formations involved, the deformation of the Plateau area is generally assumed to be Tertiary (Laramide). If deposition is related to fractures there is a strong possibility that the uranium deposits were derived from solutions originating at one side and below.

Notwithstanding the large number of uranium deposits involved, veins containing uranium minerals leading upward as possible feeders to connect with blanket deposits in strata above are conspicuously scarce. Although faults and fractures adequate to provide channels are present in many places, only limited deposition has taken place in them. This situation has caused a number of students of the origin of the Plateau uranium deposits to conclude

that hydrothermal solutions rising from below must not have participated in the formation of the blanket deposits.

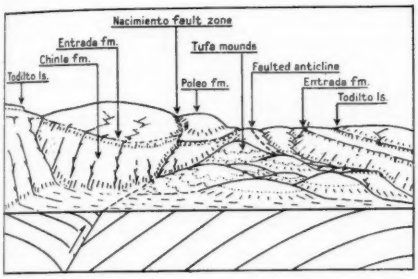

FIGURE 18.—HOT-SPRINGS DEPOSITION ALONG NACIMIENTO FAULT

Deformation accompanying faulting has provided channels for the flow of solutions. In some places along the fault the solutions are uranium-bearing. The fault lies along the eastern side of the Plateau in northeastern New Mexico.

features to the emplacement of the ores become of primary importance.

The activity of solutions along faults may be observed in the broken zone which lies close to the Nacimento fault and extends outward on either side (Fig. 18). In some places uranium deposition may be detected, although the prolific mineralization of major deposits appears related to earlier and buried solution phenomena.

At the North Alice Incline which cuts across the Lisbon Valley fault (Fig. 19) evidence of alteration is abundant, and some copper minerals are found. However, the mineralization of the Mi Vida and other uranium mines on the southwest limb of the faulted anticline has not yet been identified with the alteration so well exposed underground along the fault line.

Igneous Centers

General structure.—The penetration of Plateau strata by isolated igneous bodies sur-

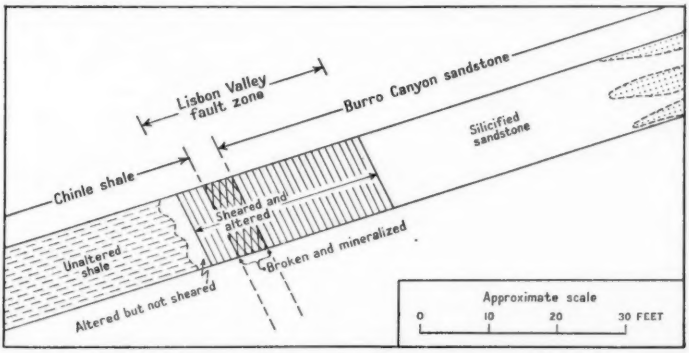

FIGURE 19.—ALTERATION ALONG LISBON VALLEY FAULT

Both argillic alteration and silicification have been effective along the Lisbon Valley fault. Copper mineralization accompanies the alteration although uranium was not observed along the fault, notwithstanding the presence of uranium mines in Shinarump strata on the south limb of the Lisbon Valley anticline (cut by the fault).

The explanation of this puzzling feature involves the origin not alone of uranium deposits in sedimentary strata, but of a number of deposits of copper, lead, zinc, and gold. However, it is well known that extensive metallic deposition may take place in more or less horizontal strata without feeder veinlets extending below. It is also a recognized feature of mineralization that the fractures along which solutions carrying metallic ions have passed are frequently not the sites in which precipitation takes place. This may have been the case in the Plateau area. If so, the abundance of structural features and the relationship of these

rounded by sediments has long been recognized (Fig. 20). Such bodies exhibit a wide range in size, form, and method of emplacement. Outstanding among the earlier Tertiary intrusives are the crystalline igneous masses of the Henry, La Sal, Abajo, Ute, Navajo, and La Plata mountains. Smaller bodies of later Tertiary intrusives consisting of dikes, sills (Fig. 21), volcanic necks (Fig. 22), cinder cones, and diatremes (Figs. 23, 24, 25) occur both within the Plateau proper and in border areas. Local vertical structures such as breccia pipes (Figs. 26, 27) and collapse features (Figs. 28, 29, 30)

connected with the hydrothermal aftereffects of igneous activity may also be observed.

The significance of the early Tertiary intrusives with respect to uranium emplacement has long constituted a problem. At Marysvale and elsewhere around the margin of the Plateau uraninite-bearing veins cut intrusives under conditions which strongly support hydrothermal origin. On the other hand, the crystalline intrusives of the mountainous masses that rise above the level of the Plateau have thus far exhibited a conspicuous scarcity of uranium veins. While a direct connection with older Tertiary intrusives has not been observed, an indirect relationship may exist. Certain belts of major uranium deposits which border these Tertiary igneous centers suggest a zonal distribution with respect to the centers.

The connection of uranium mineralization with breccia pipes, collapse features, and diatremes is more definite. Uranium may also be associated with altered areas in sediments, attributed to hydrothermal activity, in which a remote relationship to igneous activity may be inferred.

Zonal distribution.—A great belt of uranium deposits borders the La Sal Mountains on the north, east, and south in a broad arc 1–25 miles across that extends for about 125 miles. The deposits of Lisbon Valley extend the arc to the southwest, while a group of deposits nearer Moab on the west provides a further extension.

In contrast to the arc, the central area of the La Sal Mountains has been unproductive. The same sediments that carry uranium deposits a few miles from the igneous centers are both metamorphosed and barren close to the centers. If the arc of deposits is a favorable zone related to an original hydrothermal source as distribution suggests, a temperature range well below the conditions of metamorphism for sediments must apply.

The main uranium deposits of the San Rafael Swell lie a few miles east of the volcanic centers on the southwest slope of the structure (Fig. 17). The area in which dikes, sills, and volcanic craters predominate has been essentially unproductive in uranium.

The possibility of a zonal relationship also

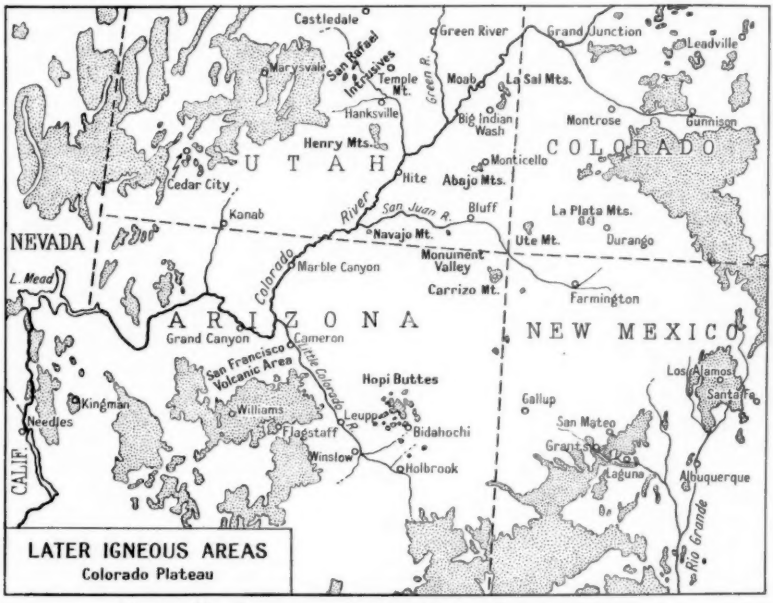

FIGURE 20.—ENCIRCLING AND PENETRATING IGNEOUS ACTIVITY

The sediments of the Colorado Plateau are encircled by areas of igneous activity and have been intruded in many places.

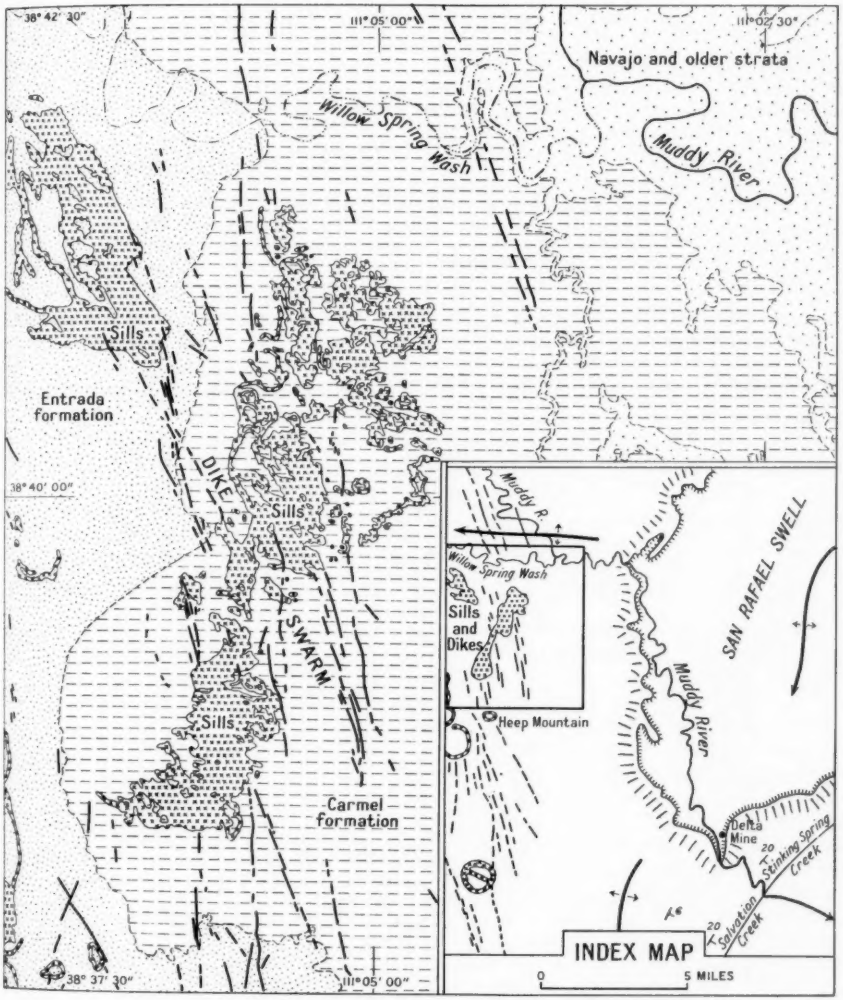

FIGURE 21.—DIKES AND SILLS IN PLATEAU SEDIMENTS

Southwest of the San Rafael Swell, Tertiary dikes and sills penetrate Jurassic strata. The volcanism is weak and produces little alteration in adjacent sediments.

exists in the vicinity of Cameron, Arizona (Fig. 31). Two collapse features—one of which is known to be mineralized with uranium—lie north of Shadow Mountain, an outlying volcanic vent on the outer margin of the San Francisco volcanic center (Pl. 3). Several such features lie south of the Little Colorado River between the main uranium area at Cameron and the Sunset Crater portion of the San Francisco volcanic area. One of these is a productive uranium deposit. The Orphan uranium deposit near the Grand Canyon Inn lies in a breccia pipe northwest of the San Francisco volcanic area. Local operators report that other

uranium-bearing pipes exist in the general vicinity.

Between Grants and Laguna, New Mexico, the Mt. Taylor volcanic area (Fig. 32) is bordered at a distance east and west by highly productive uranium areas. A breccia pipe furnished the ore body for the Woodrow Mine east of the Mt. Taylor volcanic center. Other breccia pipes are reported in the vicinity, but the Woodrow occurrence is the only pipe to have been mined.

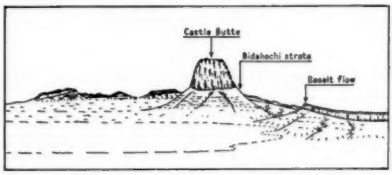

FIGURE 22.—VOLCANIC NECK IN HOPI AREA

The encircling partially eroded lava flows indicate the pre-erosional slope surrounding the vent. Similar Bidahochi strata (Pliocene) are cut by uranium-bearing diatremes.

Eruptive centers.—Tertiary volcanic centers such as those found in the San Juan, Hopi (Fig. 23), San Francisco (Fig. 31), Marysvale, and Mt. Taylor (Fig. 32) areas occur along the Plateau margin and in places merge with the Plateau. Remnants of basalt-filled volcanic necks, cinder cones, rhyolitic extrusive centers, old calderas, dikes, and diatremes are prominent in certain places. These features have long been recognized, but the possible role of this igneous activity in the emplacement of uranium deserves further attention.

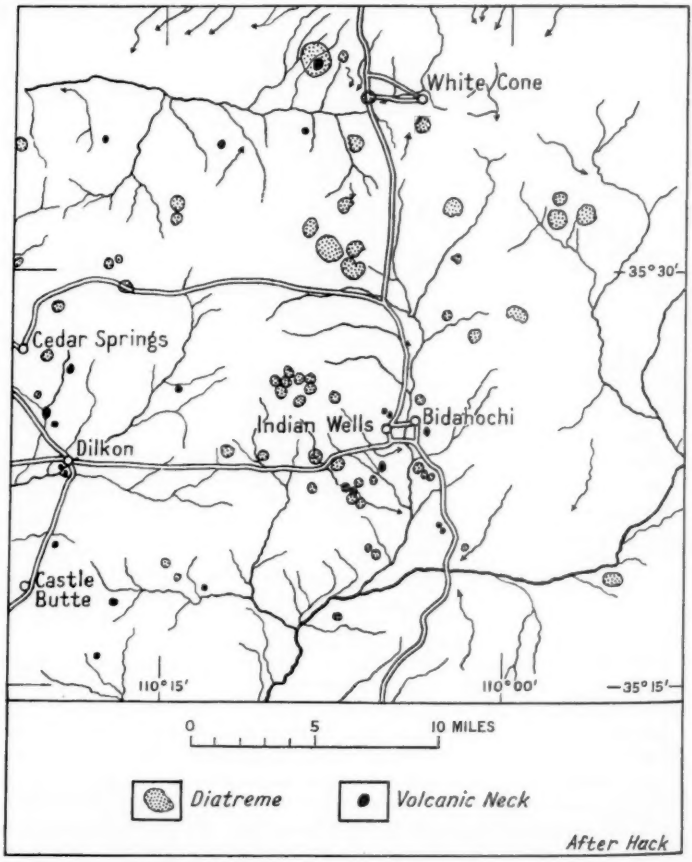

FIGURE 23.—DIATREMES OF THE HOPI AREA

Pliocene sedimentary strata in the Hopi area have been penetrated by explosive vents. The detrital remnants contain uranium in places.

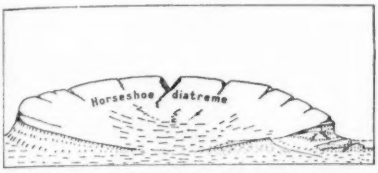

FIGURE 24.—SUNKEN DIATREME DEPRESSION
Looking north toward the Horseshoe diatreme near White Cone on the Hopi Indian Reservation, Arizona. Secondary uranium mineralization occurs in a marginal vent on the east side of the diatreme.

Even within the Tertiary these igneous features cover a considerable range in time of deposition. At Marysvale, Utah, the rhyolitic centers represent the youngest of these volcanic stages; the two preceding stages are younger than Oligocene sediments. In the Hopi area diatremes and volcanic necks penetrate sediments long considered Pliocene (Fig. 22). In the San Francisco area, Williams (1936) has recognized five stages of volcanic activity, although the part of the Tertiary represented is

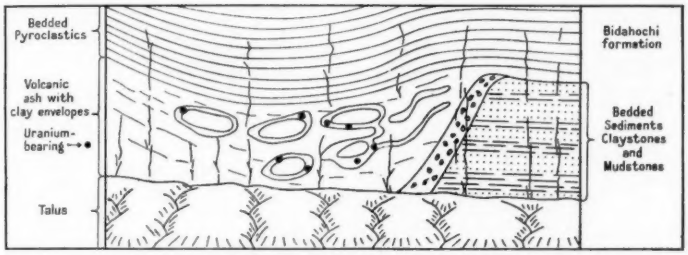

FIGURE 25.—URANIUM OCCURRENCE IN A DIATREME
Secondary uranium minerals occur in an old vent along the east side of the Horseshoe diatreme.

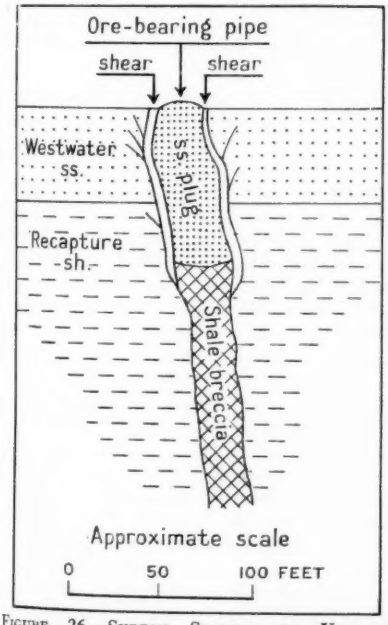

FIGURE 26.—SKETCH SECTION OF URANIUM-BEARING PIPE, WOODROW MINE
The mine is located northeast of Laguna, New Mexico, about 1 mile from the Jackpile Mine. Ore was mined from the sandstone plug for a vertical extent of about 200 feet. The pipe cuts the Westwater sandstone and the Recapture shale.

not clearly defined. Fragmentary knowledge of the portion of the Tertiary represented by these widespread igneous bodies indicates extensive late Tertiary igneous emplacement.

Uranium is associated with late Tertiary igneous activity in several places, although the extent of primary Tertiary mineralization is yet to be determined. At Marysvale, Utah, uraninite is found in a late Tertiary rhyolite dike associated with even later hydrothermal alteration. In the Hopi area, uranium is associated with late Tertiary diatremes. At Cameron, Arizona, the large sunken circular area which is probably a diatreme is mineralized with uranium (Pl. 3). Late Tertiary mineralization is suggested.

The explosion vents and magmatic plugs of these areas penetrate the sedimentary rocks. The San Francisco area lies on the southern margin of the great, generally flat-lying, sedimentary section so remarkably exposed along the Grand Canyon of the Colorado River. The Hopi area (Fig. 23) lies in the heart of the Colorado Plateau between the Painted Desert and Black Mesa. The Jemez Caldera lies along the eastern margin of the San Juan Basin. At Marysvale a great pile of volcanic rocks obscures Plateau strata other than the exposures of Deer Trail Mountain (Kerr *et al.*, 1957).

The source of the volcanism was deep-seated and was accompanied by sufficient force to

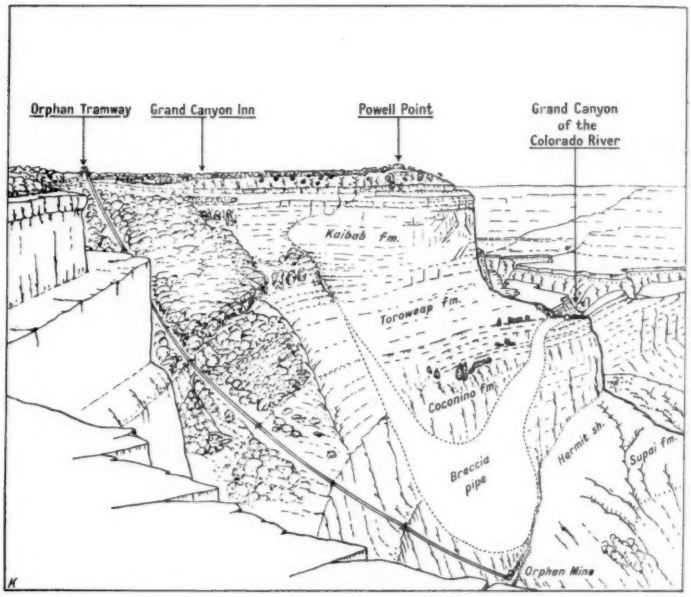

FIGURE 27.—ORPHAN MINE, GRAND CANYON, ARIZONA

A breccia pipe carrying uraninite rises vertically through the Paleozoic formations on the south side of the Grand Canyon in the vicinity of the Grand Canyon Inn. The ore introduction is accompanied by argillic alteration.

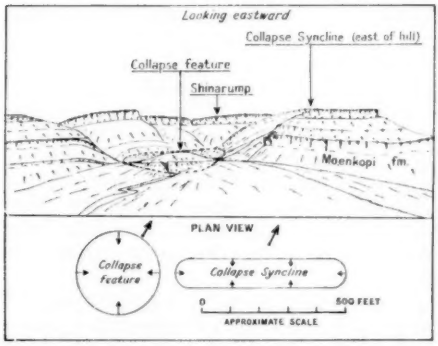

FIGURE 28.—CIRCULAR COLLAPSE AND A COLLAPSE SYNCLINE

The location is half a mile south of Arizona Highway 64, about 5.4 miles west of the intersection with U. S. Highway 89 (about 6 miles west of Cameron, Arizona).

Collapse features occur in the Moenkopi formation about 50–150 feet below the Shinarump formation. Strata show sunken forms and bleaching due to solution effects. Two features adjoin, one circular, the other long and narrow.

cause the penetration of thousands of feet of sedimentary cover which existed in late Tertiary time. In places, lava flowed out of these vents and covered many square miles of surrounding terrain. Elsewhere, explosive igneous activity scattered ash and other volcanic debris over a wide area.

Diatremes.—In the Hopi area (Fig. 23) and extending outward are numerous circular, travertine-covered elevations (Fig. 24), each with a raised rim and a sunken center. These range in diameter from a few hundred feet to about a mile. The rim normally rises several hundred feet above the Plateau, but a portion of the circumference is depressed to the level of the surrounding terrain. The travertine of the cover (Fig. 25) is uniformly stratified in layers, in many places symmetrically crenulated, and the strata slope from all sides toward the center of the structure at a moderate angle. Exposed beneath the travertine around the escarpment which encloses the structure are pyroclastic materials consisting of various forms of ejectamenta and blocks from enclosing walls. These structures are generally described as diatremes (Shoemaker, 1955). In the Hopi region they occur in the area between lava-filled vents. They were probably initiated by explosive igneous activity, but include subsequent deposition by hydrothermal solutions. They represent a feature distinct from the solid lava masses of

the vents which form the Hopi buttes, the cinder cones of the San Francisco area, or the lava-filled dikes as exemplified by Shiprock Peak in northwestern New Mexico. They appear to represent vents initiated by a volcanic explosion which created craters filled in part by volcanic debris. Subsequently, carbonate-bearing solutions emerged from the vents and deposited travertine.

Uranium association with diatremes.—The association of uranium with the diatremes (Fig. 25) is a matter of minor economic interest, but possibly of considerable significance in the interpretation of the origin of uranium on the Colorado Plateau. About 20 occurrences have been noted in the Hopi area (Shoemaker and Luedke, 1952). Secondary uranium minerals are found in veinlets and minor disseminations in tuff along the east facade of the Horseshoe diatreme. At the Seth-la-Kai diatreme, manganese and iron-impregnated layers at the base of a travertine cap yield a small amount of uranium ore (Lowell, 1956). Uranium mineralization of this type is also found in similar occurrence in another diatreme near Indian Wells.

The resemblance between the emplacement of the travertine cap on the diatremes and the formation of caps of calcareous tufa is worthy of note. At Golconda, Nevada, an extended group of tufa caps was underlain by deposits

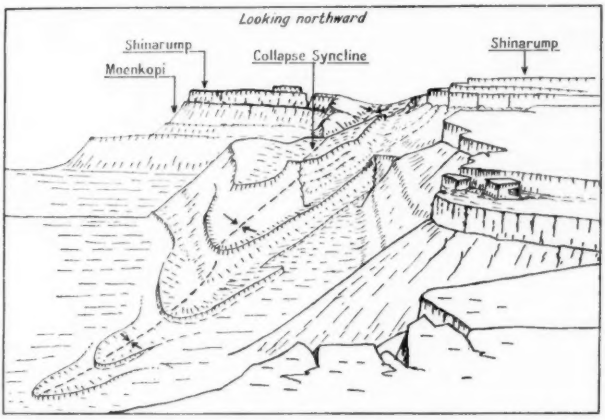

FIGURE 29.—SYNCLINAL COLLAPSE SOUTH OF CIRCULAR COLLAPSE OF FIGURE 28
Moenkopi strata show a canoe-shaped collapse tangential to a circular collapse.

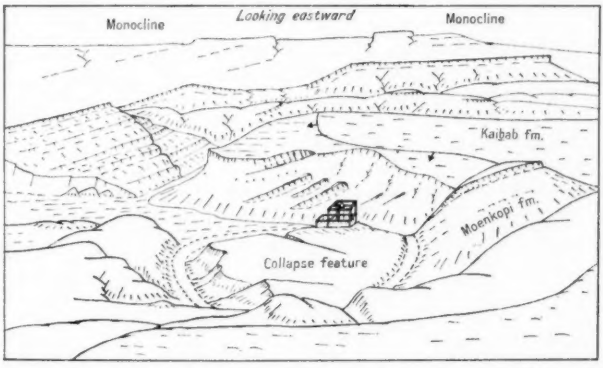

FIGURE 30.—COLLAPSE FEATURE YIELDING URANIUM, RIVER VIEW URANIUM MINE, CAMERON, ARIZONA
A small shaft has been sunk in a collapse feature in the Moenkopi formation on the edge of a monocline. River View Mine southeast of Cameron, near Black Point.

rich in manganese, iron, and tungsten (Kerr, 1940). Although tungsten has not been reported from the Hopi diatremes, the deposition of iron and manganese as hydrous oxides followed by calcium carbonate is a well-known feature of hot-spring deposits. At Monroe and Joseph, Utah, hot springs occur with large tufa caps and the customary iron-manganese rim. Strong

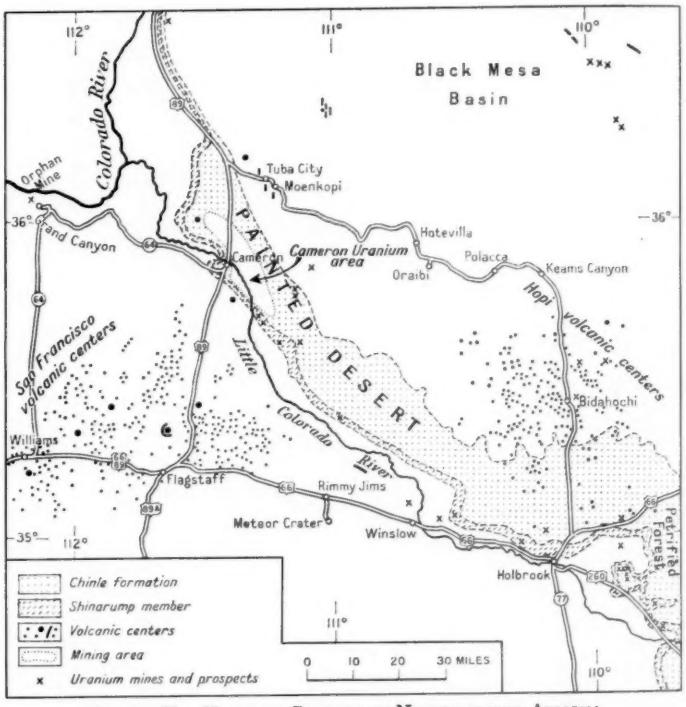

FIGURE 31.—VOLCANIC CENTERS OF NORTHEASTERN ARIZONA
The San Francisco volcanic centers are adjacent to the Cameron uranium area.

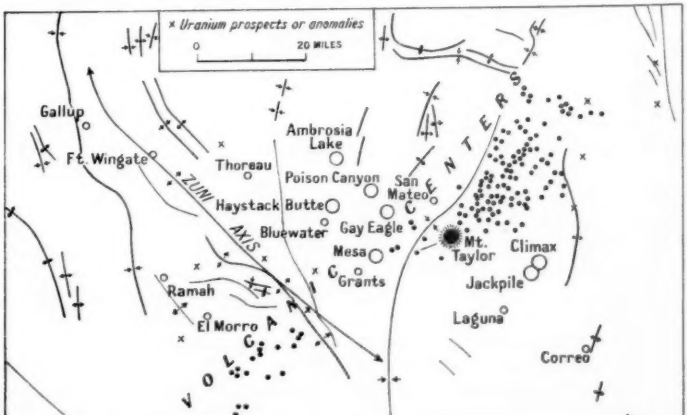

FIGURE 32.—VOLCANIC CENTERS OF THE MT. TAYLOR AREA, NEW MEXICO
A belt of volcanic centers separates the Jackpile and Ambrosia Lake uranium areas.

radioactivity is associated with the rim rather than with the tufa. In this case the radioactivity is largely due to material not in equilibrium.

The parallel with hot springs suggests that uranium-bearing solutions associated with iron pipe on the south side of the Cameron area (Fig. 30), and a collapse feature southwest of Cameron (Figs. 28, 29).

The uranium ores in the Cameron area occur in channels (Fig. 33) along which solutions may

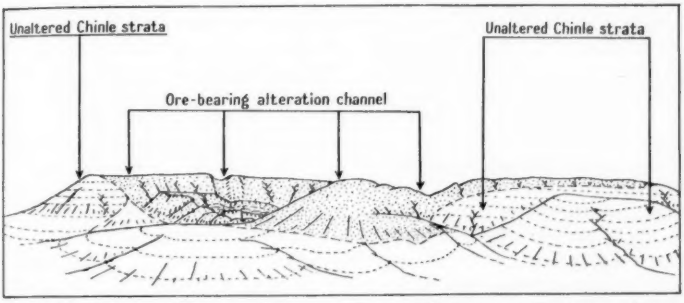

FIGURE 33.—URANIUM-BEARING CHANNEL, CHINLE FORMATION, CAMERON, ARIZONA
Yellowish-brown altered clay and sandstone unconformably overlie red-purple-gray Chinle strata which are unaltered. The main channel is about 700 feet across. The open cut of the Huskon No. 1 Mine is shown at the left. The workings are located about two miles northeast of Cameron.

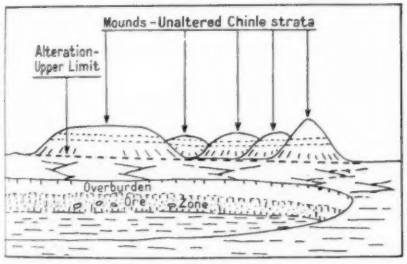

FIGURE 34.—UNALTERED CHINLE STRATA OVERLYING ALTERED-ORE ZONE
Sketch at Hoskon 7, Cameron, Arizona. Colorful strata weathered to rounded mounds overlie an ore-bearing altered channel shown in the open cut. The ore zone contains replaced logs yielding ore.

and manganese may have preceded the carbonate precipitation around diatremes in essentially the same manner that such solutions preceded the tufa caps in hot-springs areas. Since diatremes approach the uranium-bearing channels of the Painted Desert and the Cameron area, the possible role of hydrothermal solutions rising along diatremes in the emplacement of uranium ores in near-by sedimentary strata deserves consideration.

Diatreme uranium and channels.—No direct connection between a diatreme and an ore-bearing channel of the Chinle type appears as yet to have been found. The nearest approaches are probably represented by the Shadow Mountain diatreme on the outskirts of the Cameron area (Pl. 3), the Riverview uranium-bearing have migrated laterally for a considerable distance. This is indicated by the rock alteration along the channels and is associated with the ore. The channels occur at various levels in the Chinle formation (Fig. 34), and most of them are isolated, although some merge below and above.

Ores appear to represent precipitation of uranium in the vicinity of favorable types of organic detritus accumulated along channels. Presumably the solutions which carried the uranium ions deposited their load under such conditions. The localization of the ores in channels and the precipitation in an organic environment are repeatedly observed in the Cameron area. Whether the organic material itself or accompanying H_2S or both brought about precipitation is not entirely clear, but the presence of a precipitating mechanism appears established.

CONCLUSIONS

Evaluation of the criteria of emplacement leads to several significant conclusions.

Since all Plateau deposits have undergone at least some supergene action, the essential features for the interpretation of origin exist in the original mineralization. Where deposits consist entirely of supergene minerals, the original mineralization must be inferred.

The syngenetic hypothesis derives its strongest support from the wide distribution of carnotite ore in the Morrison formation. Many of these deposits consist entirely of supergene

minerals. However, where mining has continued below old water tables, primary mineralization has been found. Further, as Plateau uranium deposits have been discovered at more and more horizons, ultimately embracing a large part of the geologic column, the syngenetic hypothesis becomes less acceptable. The syngenetic hypothesis is also incompatible with a substantial portion of the absolute-age data thus far available although many additional age determinations must be made on primary material for conclusive proof.

The ash-leach theory appears adequate to explain a number of deposits. The simple leaching and deposition postulated is demonstrable, particularly in the Wyoming-Dakota area. Even here, however, some workers claim that this theory is both stratigraphically and geometrically inadequate. In many places on the Plateau, it is difficult to reconcile the ash-leach theory with field conditions. Strong doubt is also cast on this theory by the mineralization of collapse features, breccia pipes, veins in sediments, the complex mineralogy of many deposits, the elements involved, alteration features, and structural control of deposition.

The circulatory ground-water theory contributes much of value to the understanding of Plateau deposits. The extensive lateral migration of uranium ions, the relation of many deposits to the water table, and the chemistry of precipitation are all compatible with observation in many parts of the Plateau. Clearly the role of ground water must be considered. However, much evidence found fails to agree with the postulated source of uranium as derived from basement erosion. Likewise, in many places temperatures appear to have exceeded those to be expected from normal ground water.

Large areas of Precambrian rocks have been exposed several times during the erosional history of the Plateau. Even assuming that the erosion of these rocks would have provided an adequate source—which is highly doubtful—the great thickness of cover over the formations in which uranium has been found, coupled with great structural irregularity, would have precluded this origin except on a limited scale.

The evidence of alteration, the pattern of chemical elements, and the formation of uranoorganic ore all point to the prevalence in many places of temperatures above those to be attributed to ground water even where deeply buried.

The hydrothermal theory appears to explain many features which do not fit other theories. It accounts for temperatures of formation which in many instances exceed those of normal ground water. It provides wide-spread sources where conduits penetrate the basement, and solutions carrying uranium ions find their way upward along breaks to encounter permeable strata and migrate laterally through zones charged with precipitants. A well-known range in igneous activity throughout the Plateau is consistent with such absolute-age data as exist. It is consistent with the distribution of chemical elements and minerals observed. It accounts for the formation of ore bodies in collapse features, breccia pipes, and the association of uranium with diatremes, and for the alteration features associated with Plateau deposits.

On the other hand, the hydrothermal solutions can only be considered as becoming comingled with normal ground water. Also, deposits once formed have frequently become subjected to supergene solution effects with accompanying mineralogical changes.

In conclusion, it should be emphasized that data are still incomplete, yet the weight of evidence as presented in this paper favors the hydrothermal theory for the origin of Plateau uranium deposits.

In the present era of ample uranium, geologists should still keep in mind this most interesting problem. It is not only filled with geological significance, but the importance of the interpretation of Plateau origin and its connection with the long-range problem of uranium supply is a vital matter.

References Cited

Allen, E. T., Crenshaw, J. L., Johnson, J., and Larsen, E. S., 1912, The mineral sulphides of iron: Am. Jour. Sci., 4th ser., v. 33, p. 169–236

Babenroth, D. L., and Strahler, A. N., 1945, Geomorphology and structure of east Kaibab monocline, Arizona and Utah: Geol. Soc. America Bull., v. 56, p. 107–150

Baker, A. A., 1946, Geology of the Green River desert–Cataract canyon region, Emery, Wayne, and Garfield counties, Utah: U. S. Geol. Survey Bull. 951, 122 p.

Beath, O. A., 1943, Toxic vegetation growing on the Saltwash sandstone member of the Morrison formation: Jour. Botany, v. 30, p. 698–707

Beroni, E. P., McKeown, F. A., Stugard, F., and Gott, G. B., 1953, Uranium deposits of the Bulloch group of claims, Kane County, Utah: U. S. Geol. Survey TEI Circ. 239, 9 p.

Bethke, P. M., and Kerr, P. F., 1954, Uranium occurrences in the older sedimentary rocks of the Marysvale district: Annual Report for June 30, 1953, to April 1, 1954, U. S. Atomic Energy Comm., RME-3096 (pt. I), p. 60–72

Botinelly, Theodore, and Weeks, A. D., 1957, Mineralogic classification of uranium-vanad-

ium deposits of the Colorado Plateau: U. S. Geol. Survey Bull. 1074-A, p. 1–5

Botinelly, Theodore, Weeks, A. D., and Johnson, D. H., 1953, General mineralogic studies: Geologic investigations of radioactive deposits, semiannual progress report, June 1 to November 30, 1953: U. S. Geol. Survey TEI-390, p. 45–48

Breger, I. A., and Deul, M., 1955, The association of uranium with carbonaceous materials on the Colorado Plateau: U. S. Atomic Energy Comm., TEIR-539, 35 p.

Brooke, G. L., Shirley, R. F., and Swanson, M. A., 1951, Geological investigations of the trachyte district, Henry Mountains, Utah: U. S. Atomic Energy Comm., RMO-912, 7 p.

Brophy, G. P., and Kerr, P. F., 1953, Hydrous uranium molybdate in Marysvale ore: Annual Report from June 30, 1952, to April 1, 1953, U. S. Atomic Energy Comm., RME-3046, p. 45–51

Bucher, W. H., and Gilkey, A. K., 1953, Fracture pattern and uranium ore of the Zuni uplift, New Mexico (Abstract): Geol. Soc. America Bull., v. 64, p. 1402

Butler, B. S., Loughlin, G. F., Heikes, V. C., and others, 1920, The ore deposits of Utah: U. S. Geol. Survey Prof. Paper 111, 672 p.

Cannon, H. L., 1953, Geobotanical reconnaissance near Grants, New Mexico: U. S. Geol. Survey Circ. 264, 8 p.

—— 1957, Description of indicator plants and methods of botanical prospecting for uranium deposits on the Colorado Plateau: U. S. Geol. Survey Bull. 1030-M, 516 p.

Coleman, R. G., and Appleman, D. E., 1957, Umohoite from the Lucky Mc mine, Wyoming: Am. Mineralogist., v. 42, p. 657–660

Craig, L. C., Holmes, C. N., Cadigan, R. A., Freeman, R. A., Mullens, T. E., and Weir, G. W., 1951, Preliminary report on the stratigraphy of the Morrison and related formations of the Colorado Plateau region: U. S. Geol. Survey TEI-180, 64 p.

Darton, N. H., 1925, Structural features of New Mexico and Arizona: Pan-Am. Geologist, v. 43, p. 156

Derriks, J. J., and Vaes, J. F., 1956, The Shinkolobwe uranium deposit: Current status of our geological and metallogenic knowledge: Intern. Conf. on Peaceful Uses of Atomic Energy Proc., v. 6, p. 94–128

De Vergie, P. C., and Carlson, W. A., 1953, Investigation of the "C" group area, San Juan county, Utah: U. S. Atomic Energy Comm., RME-4011, 13 p.

Dodd, P. H., 1950, Happy Jack Mine, White Canyon, Utah: U. S. Atomic Energy Comm., RMO-660, 23 p.

Doelter, C., 1918, Arsen: Handbuch der Mineralchemie, Dresden und Leipzig, Verlag von Theodor Steinkopff, v. 31, p. 596–601

Erickson, R. L., Myers, A. T., and Horr, C. A., 1954, Association of uranium and other metals with crude oil, asphalt, and petroliferous rock: Am. Assoc. Petroleum Geologists Bull., v. 38, p. 2200–2218

Evans, N. N., 1903, Native arsenic from Montreal: Am. Jour. Sci., 4th ser., v. 15, p. 92–93

Everhart, D. L., 1950, Reconnaissance examinations for copper-uranium deposits west of the Colorado River: U. S. Atomic Energy Comm. RMO-659, 19 p.

Everhart, D. L., and Wright, R. J., 1953, The geologic character of typical pitchblende veins: Econ. Geology, v. 48, p. 77–96

Fenneman, N. M., 1917, Physiographic subdivisions of the United States: Nat. Acad. Sci. Pr., v. 3, p. 17–22

Finch, W. I., 1953, Geologic aspects of the resource appraisal of uranium deposits in pre-Morrison formations of the Colorado Plateau, an interim report: U. S. Geol. Survey TEI-328A, 35 p.

Fischer, R. P., 1950, Uranium-bearing sandstone deposits of the Colorado Plateau: Econ. Geol., v. 45, p. 1–11

Gilbert, G. K., 1877, Report on the geology of the Henry Mountains: U. S. Geog. Geol. Survey, 160 p.

Gilkey, A. K., 1953, Fracture pattern of the Zuni Uplift, final report: U. S. Atomic Energy Comm. RME-3050, 34 p.

Gill, J. R., 1953, Uranium minerals in Cedar Canyon, Harding County, South Dakota, *in* Search for and geology of radioactive deposits: Semiannual progress report, December 1, 1952 to May 31, 1953, U. S. Geol. Survey TEI-330, p. 124–125

Gott, G. B., Page, L. R., and Jones, R. S., 1953, Black Hills, South Dakota: Geologic investigations of radioactive deposits, Semiannual progress report, June 1 to November 30, 1953; U. S. Geol. Survey TEI-390, p. 71–79

Gray, J. R., and Tennissen, A. C., 1953, Uranium investigations near Aladdin, Crook County, Wyoming: U. S. Atomic Energy Comm. RME-4016, 13 p.

Gregory, H. E., 1950, Geology and geography of the Zion Park region, Utah and Arizona: U. S. Geol. Survey Prof. Paper 220, 200 p.

Granger, H. C., and Raup, R. B., 1956, Uranium deposits in Arizona: U. S. Geol. Survey TEI-434, 110 p.

Gruner, J. W., 1951, Origin of the uranium deposits in the Shinarump formation, a preliminary study: Ann. Report for July 1, 1950 to June 30, 1951, Pt. II: U. S. Atomic Energy Comm., RMO-837, 27 p.

—— 1956, Concentration of uranium in sediments by multiple migration-accretion: Econ. Geology, v. 51, p. 495–519

Gruner, J. W., and Gardiner, L., 1952a, Mineral associations in the uranium deposits of the Colorado Plateau and adjacent regions with special emphasis on those in the Shinarump formation: Ann. Report for July 1, 1951 to June 30, 1952, Pt. III, U. S. Atomic Energy Comm., RMO-566, 40 p.

—— 1952b, Mineral associations in the uranium deposits of the Colorado Plateau and adjacent regions, with special reference to the Shinarump formation: Sixth Progress Report for July 1, 1951 to November 30, 1951, U. S. Atomic Energy Comm., RMO-892, 23 p.

Gruner, J. W., and Knox, J. A., 1957, Minerals identified from properties in Arizona, Colorado, Montana, New Mexico, South Dakota, Texas, Utah, and Wyoming: Ann. Report for April 1, 1956 to March 31, 1957, U. S. Atomic Energy Comm. RME-3148, p. 35–51

Gruner, J. W., Gardiner, L., and Smith, D. K.,

Jr., 1953, Ann. Report, March 31, 1953, U. S. Atomic Energy Comm., RME-3044, p. 1–58
Harrison, T. S., 1927, Colorado-Utah salt domes: Am. Assoc. Petroleum Geologists Bull., v. 11, p. 111–133
Hess, Frank L., 1922, Uranium-bearing asphaltite sediments of Utah: Eng. Mining Jour., v. 114, p. 272–276
Isachsen, Y. W., Mitcham, T. W., and Wood, H. S., 1955, Age and sedimentary environments of uranium host rocks, Colorado Plateau: Econ. Geol., v. 50, p. 127–134
Kaiser, E. P., 1951, Uraniferous quartzite, Red Bluff prospect, Gila County, Arizona: U. S. Geol. Survey Circ. 137, 10 p.
Kelley, D. R., and Kerr, P. F., 1958, Urano-organic ore at Temple Mountain, Utah: Geol. Soc. America Bull., v. 69, p. 701–756
Kerr, P. F., 1940, Tungsten-bearing manganese deposit at Golconda, Nevada: Geol. Soc. America Bull., v. 51, p. 1359–1390
Kerr, P. F., and Hamilton, P. K., 1958, Chrome mica-clay, Temple Mountain, Utah: Am. Mineralogist, v. 43, p. 34–47
Kerr, P. F., and Lapham, D. M., 1954, Report on a nodule from Temple Mountain, Utah: Ann. Report for June 30, 1953 to April 1, 1954, Pt. I, U. S. Atomic Energy Comm., RME-3096, p. 7–15
Kerr, P. F., Rasor, C. A., and Hamilton, P. K., 1951, Uranium in the Black King prospect, Placerville, Colorado: Ann. Report for July 1, 1950 to June 30, 1951, U. S. Atomic Energy Comm., RMO-797, p. 25–43
Kerr, P. F., Bodine, M. W., Jr., Kelley, D. R., and Keys, W. S., 1957, Collapse features, Temple Mountain uranium area, Utah: Geol. Soc. America Bull., v. 68, p. 933–982
Kerr, P. F., Brophy, G. P., Dahl, H. M., Green, J., and Woolard, L. E., 1957, Alteration and uranium mineralization, Marysvale, Utah: Geol. Soc. America Special Paper 64, 212 p.
Keys, W. S., 1954, The Delta mine, San Rafael Swell, Emery County, Utah: U. S. Atomic Energy Comm., RME-59, 31 p.
King, D., 1953, Origin of alunite deposits at Pidinga, South Australia: Econ. Geology, v. 48, p. 689–703
King, J. W., 1955, High grade uraniferous lignites in Harding county, South Dakota: United Nations Proc. Int. Conf. Peaceful Uses of Atomic Energy, v. 6, p. 473–483
Lindberg, M. L., 1946, Occurrence of bromine in carnallite and sylvite from Utah and New Mexico: Am. Mineralogist, v. 31, p. 486–494
Love, J. D., 1952, Preliminary report on uranium in the Pumpkin Buttes area, Powder River Basin, Wyoming: U. S. Geol. Survey Circ. 176, 37 p.
—— 1953, Wyoming, reconnaissance: U. S. Geol. Survey TEI-390, p. 63–68
—— 1954, Preliminary report on uranium in the Gas Hills area, Fremont and Natrona counties, Wyoming: U. S. Geol. Survey Circ. 352, 11 p.
Lowell, J. D., 1956, Occurrence of uranium in Seth-La-Kai diatreme, Hopi buttes, Arizona: Am. Jour. Sci., v. 254, p. 404–412
McConnell, D., 1953, An American occurrence of volkonskoite: Clays and clay minerals: Second National Conf. on Clays and Clay Minerals Proc., Natl. Acad. Sci., Natl. Research Council Pub. 327, p. 152–157
Mellor, J. W., 1929, Inorganic and theoretical chemistry: London, Longmans, v. 9, 338 p.
Miller, L. J., 1955a, The chemical environment of pitchblende: Ann. Rept. for June 30, 1954 to April 1, 1955, Pt. I, U. S. Atomic Energy Comm., RME-3110, p. 1–49
—— 1955b, Uranium ore controls of the Happy Jack deposit, White Canyon, San Juan county, Utah: Econ. Geol., v. 50, p. 156–169
Moore, G. W., 1954, Extraction of uranium from aqueous solution by coal and some other materials: Econ. Geology, v. 49, p. 652–662
Moore, G. W., and Levish, M., 1955, Uranium-bearing sandstone in the White River badlands, Pennington County, South Dakota: U. S. Geol. Survey Circ. 359, 7 p.
Moulton, G. F., 1926, Some features of redbed bleaching: Am. Assoc. Petroleum Geologists Bull., v. 10, p. 304–311
Noller, C. R., 1951, Chemistry of organic compounds: Philadelphia, W. B. Saunders Co., 855 p.
Onishi, H., and Sandell, E. B., 1955, Geochemistry of arsenic: Geochimica et Cosmochimica Acta, v. 7, p. 1–33
Page, L. R., 1950, Interim report of geologic investigation, Lost Creek schroeckingerite deposits, Sweetwater County, Wyoming: U. S. Geol. Survey TEM-183A, 3 p.
Page, L. R., Stocking, H. E., and Smith, H. B., 1956, Contributions to the geology of uranium and thorium by the United States Geological Survey and Atomic Energy Commission for the United Nations International Conference on Peaceful Uses of Atomic Energy, Geneva, Switzerland 1955: U. S. Geol. Survey Prof. Paper 300, 739 p.
Palache, C., 1934, Contributions to crystallography: Claudetite; minasragrite; samsonite; native selenium; indium: Am. Mineralogist, v. 19, p. 194–205
Posnjak, E., Allen, E. T., and Merwin, H. E., 1915, The sulphides of copper: Econ. Geology, v. 10, p. 491–535
Rapaport, I., Hadfield, J. P., and Olson, R. H., 1952, Jurassic rocks of the Zuni Uplift, New Mexico: U. S. Atomic Energy Comm., RMO-642, 45 p.
Rasor, C. A., 1952, Uraninite from the Grey Dawn mine, San Juan County: Science, v. 116, p. 89–90
Schwartz, G. M., 1927, Intergrowths of chalcopyrite and cubanite, experimental proof of the origin of intergrowths and their bearing on the geologic thermometer: Econ. Geology, v. 22, p. 44–61
Seebach, M., 1926, Realgar (AsS): Handbuch der Mineralchemie Dresden und Leipzig, Verlag von Theodor Steinkopff, Doelter, C., and Leitmeir, H., Editors, v. 4 (1), p. 40–43
Shoemaker, E. M., 1955, Diatremes on the Navajo and Hopi reservations: U. S. Geol. Survey TEI-590, p. 61–71
Shoemaker, E. M., and Luedke, R. G., 1952, Map of the uranium region of the Colorado Plateau: U. S. Geol. Survey TEI-279
Smith, J. E., Jr., Hinrichs, E. N., and Luedke, R. G., 1952, Progress report on geologic studies in the Capitol Reef area, Wayne county, Utah: U. S. Geol. Survey TEI-203, 29 p.
Spence, H. S., 1931, The pitchblende and silver

discoveries at Great Bear Lake, Northwest Territories: Canada Dept. Mines, Mines Branch, Invest. Min. Res. and Min. Ind., no. 3, p. 55–92

Stewart, J. H., and Williams, G. A., 1954, Stratigraphic relations of the Triassic Shinarump conglomerate and a prominent sandstone unit of the Chinle formation in southeastern Utah (Abstract): Geol. Soc. America Bull., v. 65, p. 1387

Stieff, L. R., Stern, T. W., and Milkey, R. G., 1953, A preliminary determination of the age of some uranium ores of the Colorado Plateau by the lead-uranium method: U. S. Geol. Survey Circ. 271, 19 p.

Stieff, L. R., Stern, T. W., and Milkey, R. G., 1954, Nuclear geology: N. Y., John Wiley & Sons, Inc., 264 p.

Stieff, L. R., Stern, T. W., and Sherwood, A. M., 1956, Coffinite, a uranous silicate with hydroxyl substitution: a new mineral: Am. Mineralogist, v. 41, p. 675–688

Takeuchi, T., and Nambu, M., 1956, On the genesis of chalcopyrite lattice in bornite: Tôhaku Univ., Japan, Sci. Rep. Research Inst., A, v. 8, p. 31–44

Thompson, M. E., Ingram, B., and Gross, E. B., 1956, Abernathyite, a new uranium mineral of the metatorbernite group: Am. Mineralogist, v. 41, p. 82–103

Thompson, M. E., Roach, C., and Braddock, W., 1956, New occurrences of native selenium: Am. Mineralogist, v. 41, p. 156–157

Tourtelot, H. A., 1952, Reconnaissance for uraniferous rocks in northeastern Wind River basin, Wyoming: U. S. Geol. Survey TEM-445, 14 p.

Trites, A. F., Jr., Chew, R. T., 3d, and Lovering, T. G., 1955, Mineralogy of the uranium deposit at the Happy Jack mine, White Canyon district, San Juan County, Utah: U. S. Geol. Survey TEI-514, 48 p.

Vickers, R. C., 1953, An occurrence of autunite, Lawrence County, South Dakota: U. S. Geol. Survey Circ. 286, 5 p.

Vine, J. D., and Prichard, G. E., 1956, Geology and uranium occurrences in the Miller Hill area, Carbon County, Wyoming: U. S. Geol. Survey TEI-227, 77 p.

Warren, C. H., 1903, Native arsenic from Arizona: Am. Jour. Sci., 4th ser., v. 16, p. 337–339

Waters, A. C., and Granger, H. C., 1953, Volcanic debris in uraniferous sandstones and its possible bearing on the origin and precipitation of uranium: U. S. Geol. Survey Circ. 224, 26 p.

Weeks, A. D., Coleman, R. G., and Thompson, M. E., 1956, Summary of the mineralogy of the Colorado Plateau uranium ores: U. S. Geol. Survey TEI-583, 50 p.

Whitmore, D. R. E., Berry, L. G., and Hawley, J. E., 1946, Chrome micas: Am. Mineralogist, v. 31, p. 1–21

Willard, M. E., and Proctor, P. D., 1946, White Horse alunite deposit, Marysvale, Utah: Econ. Geology, v. 41, p. 619–643

Williams, H., 1936, Pliocene volcanoes of the Navajo-Hopi country: Geol. Soc. America Bull., v. 47, p. 111–172

Wright, R. J., 1955, Ore controls in sandstone uranium deposits of the Colorado Plateau: Econ. Geology, v. 50, p. 135–155

Wyant, D. G., 1952, Lost Creek (Wamsutter) schroeckingerite deposit, Sweetwater County, Wyoming: U. S. Geol. Survey TEM-10B, 9 p.

Yoder, H. S., Jr., and Eugster, H. P., 1955, Synthetic and natural muscovites: Geochimica and Cosmichimica Acta, v. 8, nos. 5/6, p. 225–280

Department of Geology, Columbia University, New York, N. Y.

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REBOUND IN THE BEARPAW SHALE, WESTERN CANADA

BY R. PETERSON

ABSTRACT

The Bearpaw formation (Upper Cretaceous) which occurs over large areas of Western Canada presents problems in connection with engineering works. These clay shales were consolidated under high pressures, but the load has been removed. A rebound occurred which is a maximum at the shale surface and decreases with depth. The rebound is made up of elastic rebound which occurred immediately upon removal of load and a "time rebound" which is still occurring slowly and involves a softening of the material.

The characteristics of the shale and the depth of the rebound zone are being studied by soil-mechanics methods. Routine tests were made for water content, density, and Atterberg limits; water content is the best indicator of consistency. The results of laboratory strength tests do not correlate well with observed field behavior, and this has led to the conclusion that observations of existing slopes will supply more reliable strength data for long-time stability studies. Field investigations involve piezometric measurements and observations of slides, particularly creep. The softened surface zone is unstable, and slides have occurred on flat slopes. Slickensides are common in this zone, but their intensity appears to decrease with depth. The underlying shale is firm and stable.

Rebound is also caused by local excavations for structures. Observations are being taken on light concrete structures to correlate predictions from laboratory swelling tests with observed heave.

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INTRODUCTION

The Bearpaw shale and other clay shales crop out or occur beneath the overburden over large areas of Western Canada. These shales have been heavily preconsolidated by sediments and ice and are now in a state of rebound or expansion because of the release of load. The movement associated with the rebound or expansion is vertical and also horizontal adjacent to the valleys. These movements result in serious problems in the design of engineering works and exert a major influence on slope stability.

The rebound can be divided into two phases: (1) elastic rebound occurs immediately upon release of load and varies directly with the load; (2) "time rebound" takes place over

many thousands of years. "Time rebound" involves swelling and an increase in water content, which is the reverse of consolidation, and further volume increase, caused by rearrangement and expansion of particles, known as the secondary time effect. This paper is concerned principally with "time rebound".

The term clay shale as used herein refers to an uncemented sediment that consists largely of clay-sized particles and exhibits plasticity and other clay characteristics when manipulated in the presence of water.

Figure 1 shows the extent of the Bearpaw shale and other clay shales in the prairie area of Western Canada and in the adjacent areas of Montana and North Dakota. The Canadian Government sponsored extensive investigations for a major multipurpose dam on the South Saskatchewan River in Saskatchewan at the location shown by the solid square on Figure 1. The foundation and abutments at this site consist of Bearpaw shale, and most of the information presented herewith will deal with studies at this location. In addition to extensive laboratory testing on samples recovered from test holes, a number of shafts and a drift have been utilized to obtain undisturbed samples and observe the material in place. Observations are being taken to record creep and piezometric pressure.

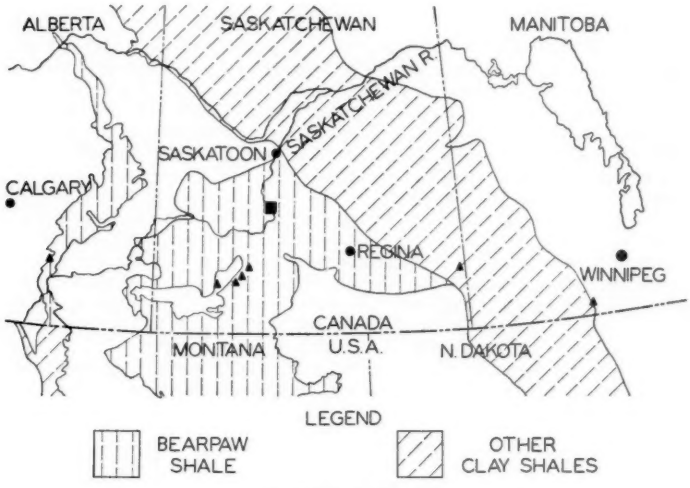

FIGURE 1.—CLAY-SHALE AREAS
Solid triangles denote dams where rebound measurements are being taken.

ACKNOWLEDGMENTS

Many people have contributed to the studies on the Bearpaw shale which have been carried out and are still being carried on by the Prairie Farm Rehabilitation Branch of the Canadian Department of Agriculture. The author is especially grateful to Profs. K. Terzaghi and A. Casagrande for their helpful advice on the over-all study and for their comments on this paper. Profs. J. B. Mawdsley, F. H. Edmunds, A. R. Byers, and W. O. Kupsch of the Geology Department of the University of Saskatchewan also made valuable suggestions and provided details regarding the local geology. Messrs. G. Parkinson and C. Booy of the Prairie Farm Rehabilitation Administration Design Office have suggested worthwhile improvements. Several of the author's colleagues including A. S. Ringheim, P. J. Rivard, and J. L. Jaspar assisted in compiling the data and preparing the figures.

GEOLOGY OF THE AREA STUDIED

The clay shales of Western Canada are of marine origin and Late Cretaceous age (Edmunds, 1944, p. 229). These sediments were deposited in the Pierre Sea which extended from the Arctic to the Gulf of Mexico. They

are generally uniform and medium gray. They are soft, slake easily with water, and on moistening expand to form a highly plastic clay. These shales contain montmorillonite, and in some places layers of bentonite are present.

At the proposed South Saskatchewan River dam site approximately 2000 feet of sediment had been eroded before the first glaciation. The geology is very complex, but it is believed that the area may have been glaciated three times with active downcutting during the interglacial stages (Mollard, 1955, p. 3). In the post-glacial period, before formation of the present stream, deposition took place in lakes, and several stages involved large fluctuations in water levels.

Description of Shale at South Saskatchewan River Dam Site

General

The Bearpaw shale at the South Saskatchewan River dam site is soft and weathered at the surface but increases in strength and density with depth. The soft surface material is erratic in consistency and can be easily remolded under finger pressure. It contains slickensides and joints which are generally closely spaced, and many surfaces are glossy and wet. In many instances there are signs of movement and sliding within this zone. The underlying undisturbed material is harder and is generally dull in appearance but becomes shiny if rubbed with a knife or the fingernail. Chunks can be broken in the hands only with difficulty, and the material becomes harder with depth. It is more nearly uniform in the hard zone, and the number of slickensides and joints decreases with depth. Weathering is generally limited to the surface zone and is characterized by a brown coating on the joints. There is only limited evidence of any layers containing pure bentonite or any other marker beds in the formation explored at this site. However, some of the layers are more plastic and are referred to as bentonitic shale.

Test Results

Complete test results on this shale are given elsewhere (Peterson, 1954, p. 19). Some of the terms used in describing the shale are defined as follows:

The void ratio e is defined as the total volume of voids divided by the volume of the solid particles; where the voids are filled with water, the void ratio becomes the volume of water divided by the volume of solids.

The water content w is given in per cent and is the weight of water divided by the weight of the solids as determined by oven drying. This is in contrast to the definition of water content as used in geology where the basis is total weight of sample rather than weight of soil solids. By substituting as shown in Figure 2 the water content can be expressed in terms of void ratio and specific gravity of solids. The void ratio e is equal to the specific gravity of the solids G times the water content w.

The volume change is equal to the change in void ratio Δe divided by 1 plus the original void ratio. Substituting the value of void ratio in terms of water content into this expression then gives the volume change in terms of the change in water content, the original water content, and the specific gravity of soil solids. For the shale under consideration the average specific gravity of the solids is 2.77; therefore the volume change equals $2.77\,\Delta w/1 + 2.77w$, and the volume change can be computed when the initial water content and change in water content are known.

The Atterberg limits referred to are the liquid limit and the plastic limit. The plastic limit is the minimum moisture content necessary to change a soil from the semisolid stage to the plastic stage; soil is considered plastic if it can be rolled into threads with a diameter of about an eighth of an inch. The liquid limit is the moisture content necessary to change the soil from the plastic stage to the semiliquid stage; this value is determined by special apparatus.

The shale consists predominantly of clay sizes less than .005 mm; the average grain-size determination indicates approximately 50 per cent of this size range—about 33 per cent silt sizes and the balance fine-sand sizes. The liquid limit ranges from 80 to 150, and the plastic limit from 18 to 27 for typical shale samples. For the bentonitic shale, however, the liquid limit ranges up to 250, and the plastic limit to 47. The moisture content of the wetter parts of the soft zone is of the order of 35 per cent and about 19 per cent in the deeper portions of the hard zone. The corresponding dry densities are about 85 pounds per cubic foot for the soft zone and range to 108 pounds per cubic foot in the hard zone. The unconfined compressive strength ranges from a minimum of 7 pounds per square inch for the softest material to 400 pounds per square inch for the hard material.

Index Tests

The water content is the most suitable test to indicate consistency and strength of the shale. Near the ground surface where drying has occurred the water content is considerably lower than in the underlying material. Below the zone of drying the moisture content generally shows a decrease with depth. The moisture content determined at 6-inch intervals and Atterberg limits for a typical test hole are shown in Figure 3.

Consistency Zones

In order to study the variations within the shale three consistency zones were arbitrarily established. These were based largely on water content and somewhat on density and shear strength. The soft surface zone was considered to have a moisture content in excess of 29 per cent, the medium zone from 25 to 31 per cent, and the underlying hard zone less than 27 per cent. With these approximate values and corresponding density and strength limita-

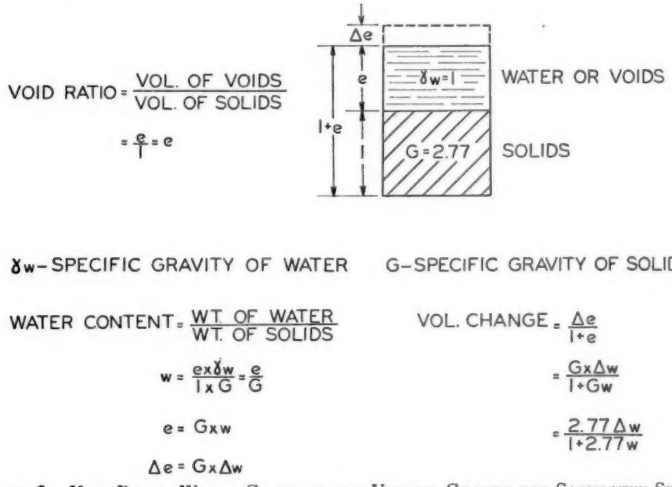
VOID RATIO = VOL. OF VOIDS / VOL. OF SOLIDS
= e/1 = e
Δe
e
1+e
γw=1
WATER OR VOIDS
G=2.77
SOLIDS
γw-SPECIFIC GRAVITY OF WATER
G-SPECIFIC GRAVITY OF SOLIDS
WATER CONTENT = WT. OF WATER / WT. OF SOLIDS
w = e x γw / 1 x G = e/G
e = G x w
Δe = G x Δw
VOL. CHANGE = Δe/1+e
= G x Δw / 1+Gw
= 2.77 Δw / 1+2.77w

FIGURE 2.—VOID RATIO, WATER CONTENT, AND VOLUME CHANGE FOR SATURATED SEDIMENT

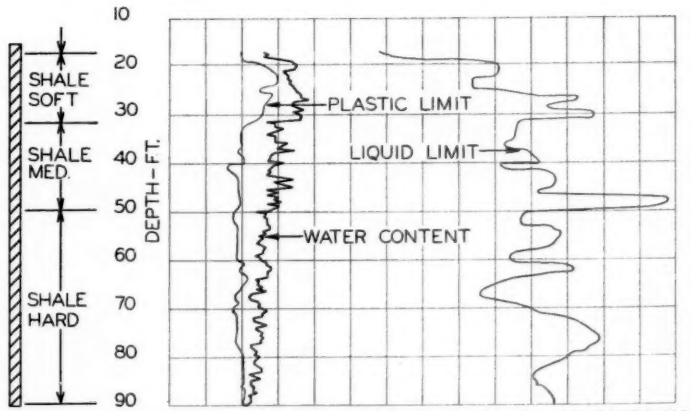

FIGURE 3.—TYPICAL TEST-HOLE RESULTS WITH SHALE ZONES

Below the zone of drying the moisture content decreases with depth and approaches the plastic limit at the bottom of the bore hole. The consistency zones are shown at the left.

tions the test results were studied and the zone boundaries established in the shale section of all test holes.

In order to study the occurrence and thickness of the consistency zones numerous profiles

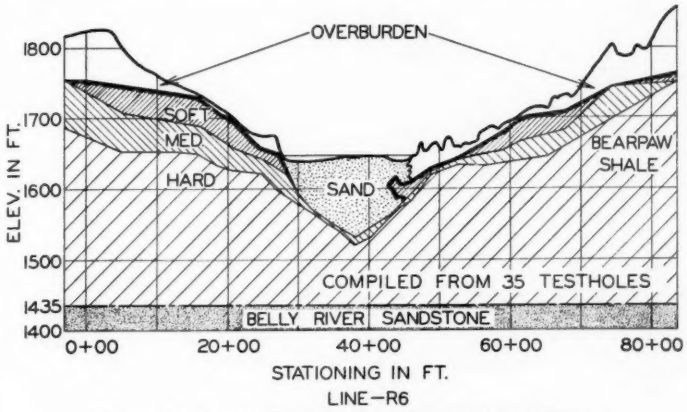

FIGURE 4.—TYPICAL RIVER PROFILE SHOWING SHALE ZONES

have been drawn. Contours representing the surface of the soft, medium, and hard zones have been constructed for the site area. Figure 4 is a typical profile across the river valley showing the consistency zones. The maximum thickness of the combined soft and medium zone occurs on the valley slopes, whereas the minimum thickness occurs below the sand in the river bed and on the upland. However, the buried slopes were once located above the valley floor, and probably most of the softened shale on these slopes has been removed by erosion. Possibly the lower levels of the valley were exposed for a shorter period than the upper portions, and therefore the softening might not have progressed to the same depth. The unusual tongue or projection about mid-height in the river sand on the right side of this profile has been located in a number of test holes. It consists of soft shale and is probably an overthrust caused by sliding or lateral swelling or both. Many of these slopes show old scars that have formed as a result of lateral creep or sliding toward the river channel.

Slide Topography

Examination of the banks in the vicinity of the dam site indicated the presence of what appear to be extensive slides, although active movement is not evident at present except possibly where erosion is taking place because of river action at the toes of slopes in critical areas. Laboratory strength tests on the soft shale give shear strengths several times greater than those computed from an ordinary circular arc analysis in a slide area, assuming a factor of safety of unity. This assumes that a slide has occurred and makes no allowance for the lateral forces that may have been at work in producing these movements. Terzaghi (1955, p. 602), in discussing this landslide topography, refers to the energetic lateral expansion of the shale on both sides of the valley toward the trough formed by erosion. He states that the movement in the shale probably assumed the character of a gravity creep, which resulted in the formation of ridges and troughs in the overlying glacial overburden. Observations are now being made to check the nature of the movement and determine whether equilibrium has been reached. A number of vertical plastic pipes have been installed in the river slopes, and lateral movement is being recorded by the use of a slope indicator developed by Stanley Wilson of Seattle.

Plate 1 shows typical photographs of the scarred river banks.

Slickensides

Slickensides were noted in the earliest investigation at this site. Many geologists have speculated on whether or not they were related to active slides. Casagrande (1949, p. 15) pointed out that many clay shales contain numerous, irregular slickensides, and it is usually assumed that such slickensides developed during consolidation of the material.

He considers it more likely that most of the slickensides formed during swelling or rebound, which resulted from the decrease of the maximum overburden pressure to the existing values. His principal reasons for this are:

(1) After the material is fully consolidated it is more sensitive to development of shear failure from irregular strains.

(2) Irregular strains are more likely to develop during unloading, since that process produces much less regular stress conditions than those prevailing during deposition of additional load.

Casagrande states that if this hypothesis is correct, it would follow that highly slickensided clay shales have greater swelling tendencies than those clay shales having only very few or no slickensides, and he found this to be the case in a number of instances.

The information obtained in this study seems to support Casagrande's hypothesis, as the slickensiding seems to be more intense in areas of greatest rebound and decreases with depth.

Rebound Estimates

General

Moisture content is greatest near the surface of the shale and decreases with depth. The shale can be divided into several consistency zones; the softest material occurs at the surface. Presumably at one time before erosion the shale mass at the site elevation was of the same consistency and as hard or harder than the zone now classed as hard shale. The vertical pressure due to the overlying sediments and glacial ice probably reached a maximum value of 100–150 tons per square foot. As material was eroded apparently the elastic rebound took place. In the time interval since erosion a "time rebound" has been going on which has involved an increase in moisture content and decrease in density; maximum softening occurs at the shale surface. In view of the fact that the material is saturated and consists of solid particles and water the water content is an indicator of the volume. In subsequent considerations volume changes will be computed from changes in water content.

The surface topography indicates horizontal expansion and/or sliding in the shale, and there is also evidence of high horizontal pressures within the shale. In the excavation of a test drift about 300 feet long the soft shale in particular moved laterally into the opening. The total inward movement of both sides amounted to as much as 20 inches in a drift 6 feet wide. The corresponding inward movement within the hard shale was of the order of 2 or 3 inches; upward heave of the floor was less than three quarters of an inch. The horizontal and vertical pressure within the hard shale was measured by a special test section (Peterson, 1954, p. 7). These observations indicate that the lateral pressure within the hard shale is approximately 150 per cent of the vertical pressure at a point where the combined thickness of overburden and shale is 65 feet. The vertical pressure at this point is approximately equal to the existing overburden pressure.

Consolidation Tests

The consolidation test is used to study the pressure-volume relationship of soils or soft rocks. This test involves placing an undisturbed sample in a confining ring which is generally about 3 or 4 inches in internal diameter and approximately 1 inch in height. Provision is made to apply known pressure increments to this disc-shaped sample and to measure the amount of compression that takes place. Figure 5 shows a typical laboratory consolidation curve for hard shale. The volume or the sample thickness is indicated on the ordinate by both void ratio and water content. The pressure is plotted on the abscissa to a logarithmic scale. *A–B* represents the loading portion of the curve, and *B–C* represents the unloading or rebound portion determined as the load is reduced from the maximum value of 50 tons per square foot. The solid line is that portion determined in the test, whereas the

Plate 1.—TYPICAL RIVER-BANK TOPOGRAPHY, SOUTH SASKATCHEWAN RIVER

Figure 1.—View looking upstream with the river at the left and the upland at the right. The overburden consisting principally of sand is of the order of 80–100 feet below the upland and becomes thinner toward the river. Movement in this area appears to have been largely horizontal; lateral movement of at least 200 feet took place, *i.e.*, from point *A* to point *B*.

Figure 2.—Looking downstream with the upland at the left and the river at the right. Characteristic ridges and troughs parallel the river. Movement was largely horizontal.

FIGURE 1

FIGURE 2

TYPICAL RIVER-BANK TOPOGRAPHY, SOUTH SASKATCHEWAN RIVER

dotted portion from *B* to *D* has been projected for the purpose of estimating the void ratio or water content corresponding to the maximum estimated pressure of 150 tons per square foot. In view of the fact that the Bearpaw geological load. Point *D* on the laboratory curve produced should provide an estimate of the void ratio and water content that existed in the formation under the maximum estimated load of 150 tons per square foot. At point *D*

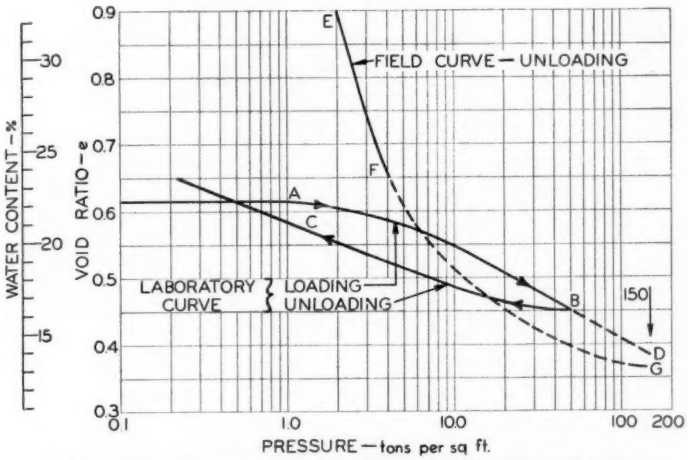

FIGURE 5.—LABORATORY AND FIELD PRESSURE-VOID RATIO CURVES

formation at the site is now in the unloading stage one would expect that the conditions in the ground might be represented by the rebound portion of the curve *B–C*. In order to check this Casagrande (1949, Fig. 9) plotted the void ratio-pressure curve *E–F* that exists at the dam site based on the void ratio of samples and the estimated pressure at the elevation of each sample. He also estimated the relation that might have occurred in the field under greater loads, that is, *F–G*. More recently the curve *E–F* has been confirmed by additional and deeper borings. For the same pressure range the field curve *E–F* is much steeper than the laboratory rebound curve *B–C*. In other words, much more swelling or expansion has taken place in the field than occurred in laboratory tests. This is probably due to the creep or secondary time effect which has taken place in the field over generations, compared with the small effect determined in the laboratory test where each load increment is applied for only 1 or 2 days. Therefore the laboratory consolidation and rebound test probably cannot be used to reconstruct a field rebound curve. However, it can be used to provide some clues to the range of moisture content or void ratio that existed in the shale formation when it was subjected to maximum the moisture content is about 14 per cent, whereas Casagrande's estimated curve at *G* gave a value of 13 per cent—a relatively good check.

The slope of the laboratory rebound curve is known as the swelling index, and it is the change of void ratio corresponding to a tenfold change in pressure. This slope for most laboratory tests was nearly constant and averaged .06 for the hard-shale samples. In the pressure range *E–F* the field curve is more than 10 times steeper than the laboratory rebound curve.

Time Rebound Based on Water-Content Change

Figure 6 shows the left half of the river-valley profile shown in Figure 4. With a view to making an estimate of the amount of "time rebound" that has occurred the average water content-depth profiles in several areas were determined. One group of five holes was in the upland area in the vicinity of *A*, and the averaged results have been plotted on the water-content chart at the left as line *A–D*. Similarly water content-depth profiles were determined for locations *E* and *H*, and they also have been shown on the moisture-content plot as line *E–G* and *H–I*. On all three profiles

the average moisture content at the surface of the shale is about 30 per cent, and it decreases with depth. At point *D* it decreases to 19 per cent and at point *I* to 22 per cent.

Before computations can be made of the amount of rebound it is necessary to estimate the moisture content that existed in the shale under maximum geological load and also to apportion the rebound between the amount that occurred in a vertical direction and the amount that occurred in a horizontal direction. The water-content profiles (Fig. 6) indicate that the water content was probably less than 19 per cent under conditions of maximum geological load. Examination of typical consolidation curves projected to a pressure of 150 tons per square foot indicates that the moisture content of the consolidated shale would probably have been about 14 per cent; this checks with Casagrande's estimate. As no method for estimating the relative amounts of vertical and horizontal rebound has been found, the two extremes, that is, rebound only in the vertical direction and rebound only in the horizontal direction toward the valley, have been considered.

Assuming that all the swelling or "time rebound" occurred in a vertical direction and that the moisture content of the shale increased from an initial moisture content of 14 per cent to the values given by the curves *A–D*, *E–G*, and *H–I*, calculations were made of the vertical heave. On this basis point *A* heaved or increased in elevation with respect to point *D* by approximately 53 feet. Similarly it was computed that point *E* had risen 36 feet and point *H* 26 feet. As most of this movement occurred in the upper portion of the shale it is apparent that differential movements occurred with respect to horizontal planes. Assuming that all the rebound occurred in a horizontal direction and that the moisture content increased from the original estimated value of 14 per cent to the present water content between *B* and *E*, the horizontal rebound or movement of point *E* away from point *B* was computed to be about 600 feet.

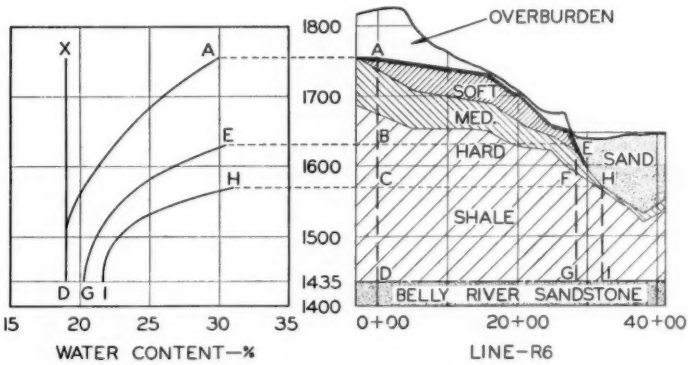

FIGURE 6.—AVERAGE WATER-CONTENT DEPTH PROFILES

If the original moisture content in the shale was 19 per cent rather than 14 per cent the above values would be reduced by approximately a half.

The foregoing computations of "time rebound" are based on the assumption that either all the movement or volume change took place in the vertical direction or all movement took place in the horizontal direction. The rebound took place in both directions; therefore these values would be reduced accordingly, but no method has been devised to estimate the portion in each plane. The writer hoped that a comparison of water-content profiles from the valley slopes, where both vertical and horizontal rebound occurred, with profiles in the bottom of the valley and on the upland, where little or no horizontal rebound was thought to have occurred, would provide useful data. However, a preliminary review of this information has not yet led to any conclusions. One reason for this is that there are insufficient holes of adequate depth in the shale on upland areas. Perhaps this approach cannot be utilized because of the complex geological history and variations in material and unloading conditions.

Application to Existing Structures

Most of the studies were carried out on the Bearpaw shale at the site of the proposed multipurpose dam on the South Saskatchewan River. However, an attempt has been made to apply these results to the design and construction of concrete chute spillways located on other clay shales. The main difficulty in the design of these structures is heave of the relatively light floor slabs. The heave, which is due to elastic rebound caused by the excavation, takes place immediately and does not present a problem. However, the upward movement due to the "time rebound" as a result of excavation does present a serious problem.

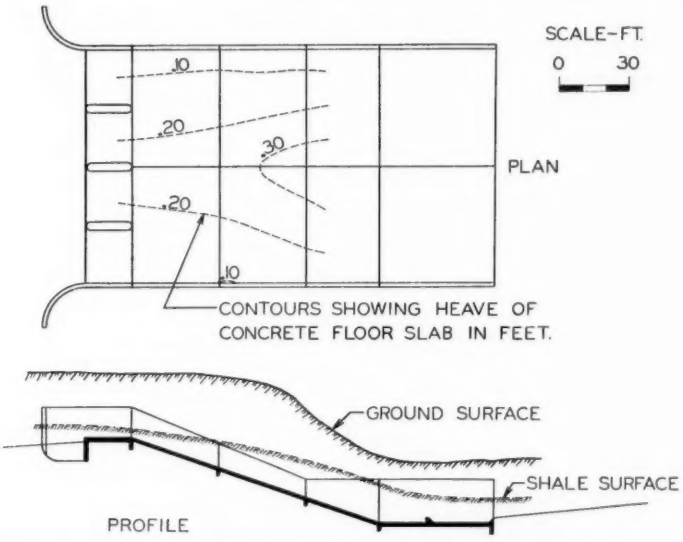

Figure 7.—Heave Contours of Spillway Floor Slab

Figure 7 includes a plan and profile of a typical spillway structure. Contours showing the heave of the concrete floor slab are indicated by the dashed lines on the plan view. The maximum heave is in excess of .30 feet, and this occurs near the center of the slab and at the point of maximum excavation. Observations indicate that the rebound is still taking place but at a decreasing rate.

In an attempt to estimate the heaving tendencies of the different clay shales a program involving laboratory testing is being carried out using equipment similar to that required for the consolidation test. Undisturbed samples are placed in the confining ring, and the amount of swell or rebound is determined upon saturation at various unit pressures. The test tentatively adopted as a standard after studying methods in use by the U. S. Corps of Engineers (1954, p. 13) and the Bureau of Reclamation (Holtz and Gibbs, 1954, p. 6) involves applying loads to the sample and then saturating the material. Relatively light loads in the range of 1 psi to 20 psi are generally used, and the amount of swelling under a definite load is observed over a period of time.

In addition to testing samples in the undisturbed state at natural moisture content tests are also being carried out on dried undisturbed samples to determine the effect of field drying on the swelling characteristics. Distilled water is generally used to saturate the soils, but other waters and "shale juice" are also used. Figure 8 shows typical results for this type of swelling test. For both dried and undried specimens greater swelling occurs at the lower pressure. Samples that have been dried out before saturation exhibit much greater volume increase than samples tested at natural water content.

Sufficient data are not yet available to correlate results of field observations with laboratory tests, but some general conclusions are evident. As would be expected there seems to be less tendency for "time rebound" to

occur where the structures are founded at shallow depths on weathered or soft shale, compared to those founded on the harder shale at greater depth. Surface drying

structures in the soft shale. In several case piles cast in place and belled at the botton have been used successfully to reduce rebounc The shale surface is in many cases protecte

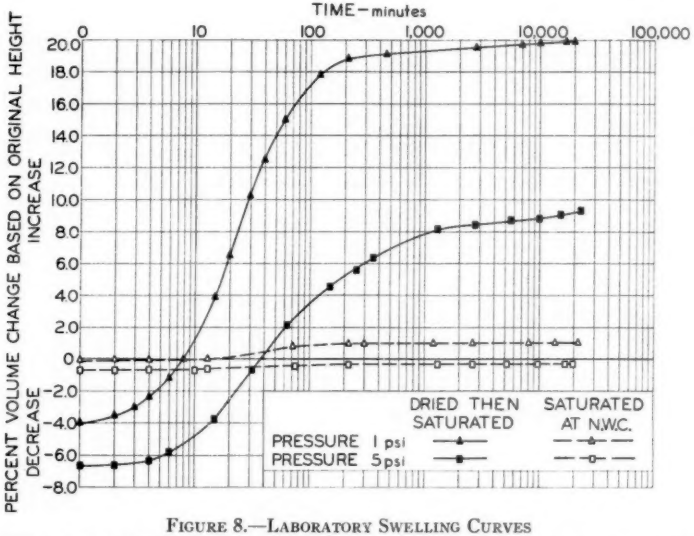

FIGURE 8.—LABORATORY SWELLING CURVES

The per cent volume change is indicated as ordinate; increase in volume is above the zero line and decrease in volume below the zero line. The abscissa is a log plot showing time in minutes. The solid points and the full lines represent tests on clay shales that have been dried before saturation. The open points and the dotted lines represent tests on samples that have been saturated at their natural moisture content. The triangular points are for a pressure of 1 psi and the square points for pressure of 5 psi.

apparently increases the tendency to swell when the material is subsequently saturated.

CONCLUDING COMMENTS

The Bearpaw shale that occurs at the South Saskatchewan dam site is probably among the most expansive yet encountered. The expansion or rebound that has occurred is indicated by a high water content at the softened shale surface and a pronounced decrease of water content with depth. Possibly the rebound is now nearly completed, and relatively stable conditions exist in the natural material. However, the design of engineering works requires very careful consideration of deep excavations and problems relating to slope stability. Deep excavations or unloading will result in additional rebound. In view of this the tentative approach is to locate structures as shallow as possible and to keep the slopes very flat. In some instances this makes it necessary to found to reduce evaporation and drying and thus avoid swelling upon resaturation. Alternatives that involve loading rather than unloading may be more desirable from the stability point of view.

The problems involving the shale and rebound are similar to those being encountered by the U. S. Corps of Engineers (Smith, 1953, p. 62; Lane and Occhipinti, 1953, p. 402) at the dams under construction on the Missouri River.

REFERENCES CITED

Casagrande, A., 1949, Notes on swelling characteristics of clay shales: Cambridge, Mass., Harvard Univ., 16 p.

Corps of Engineers, U.S. Army, 1954, Report of investigation of expansive characteristics of shale and weak rocks (Draft), Missouri River Div., Omaha, Neb., 63 p.

Edmunds, F. H., 1944, Geology and its relationship to soils in Saskatchewan: Univ. Saskatchewan, Dept. Geology, 31 p.

Holtz, W. G., and Gibbs, H. J., 1954, Engineering properties of expansive clays: Am. Soc. Civil Engineers Proc., v. 80, no. 516, 28 p.

Lane, K. S., and Occhipinti, S. J., 1953, Rebound gages check movement analysis at Garrison Dam: 3d Internat. Conf. on Soil Mechanics and Found. Eng. Proc., v. I, p. 402–405

Mollard, J. D., 1955, South Saskatchewan River project summary report on geology site 10: Regina, Saskatchewan, Air Photo Analysis and Eng. Geology Div., Eng. Services, Prairie Farm Rehabilitation Adm., 38 p.

Peterson, R., 1954, Studies of Bearpaw shale at a damsite in Saskatchewan: Am. Soc. Civil Engineers Proc., v. 80, no. 476, 28 p.

Smith, C. K., 1953, Soil properties of Fort Union clay shale: 3d Internat. Conf. on Soil Mechanics and Found. Eng. Proc., v. I, p. 62–66

Terzaghi, K., 1955, Influence of geological factors on the engineering properties of sediments: Harvard Soil Mechanics Ser. 50, p. 602–603

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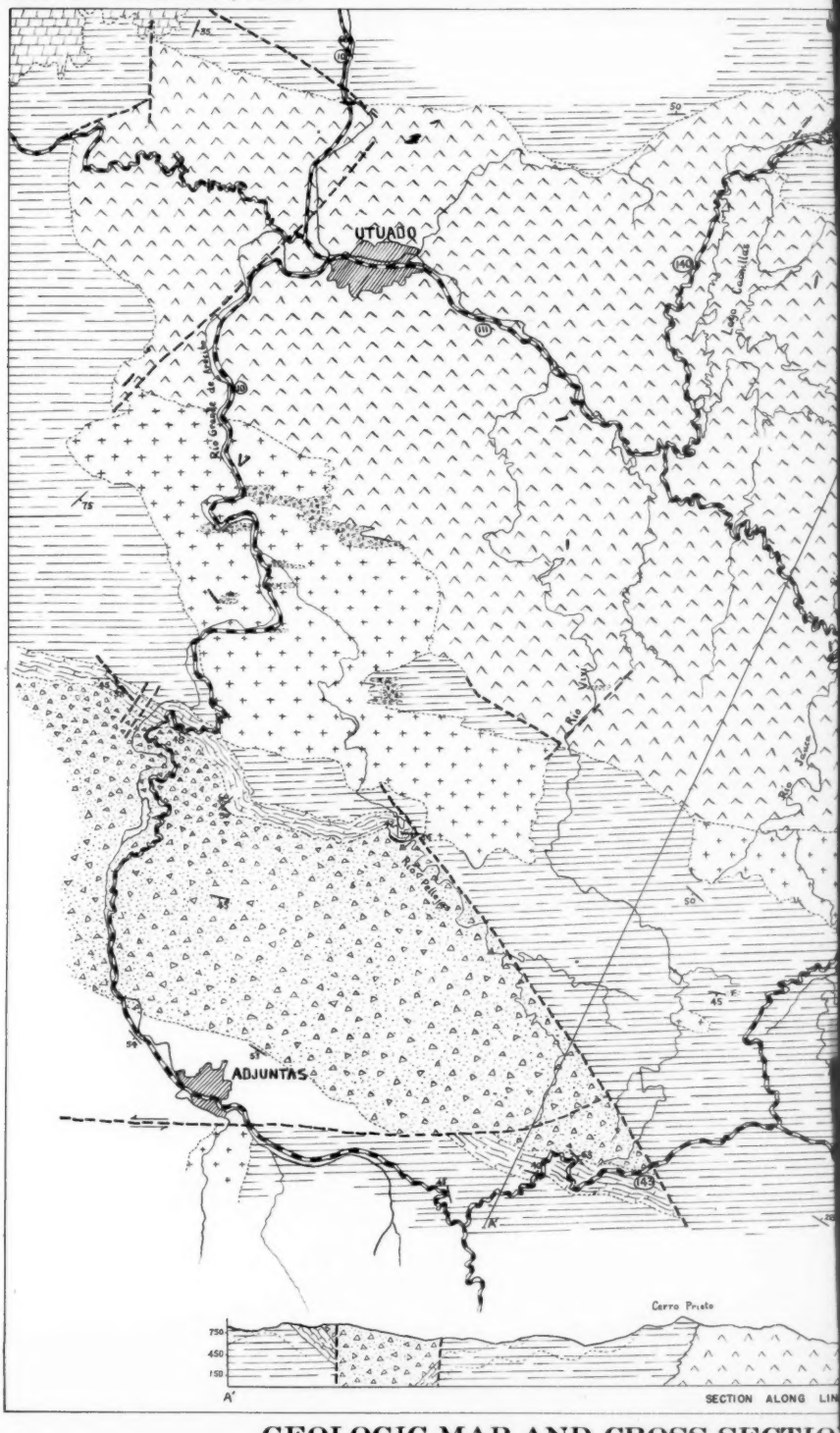

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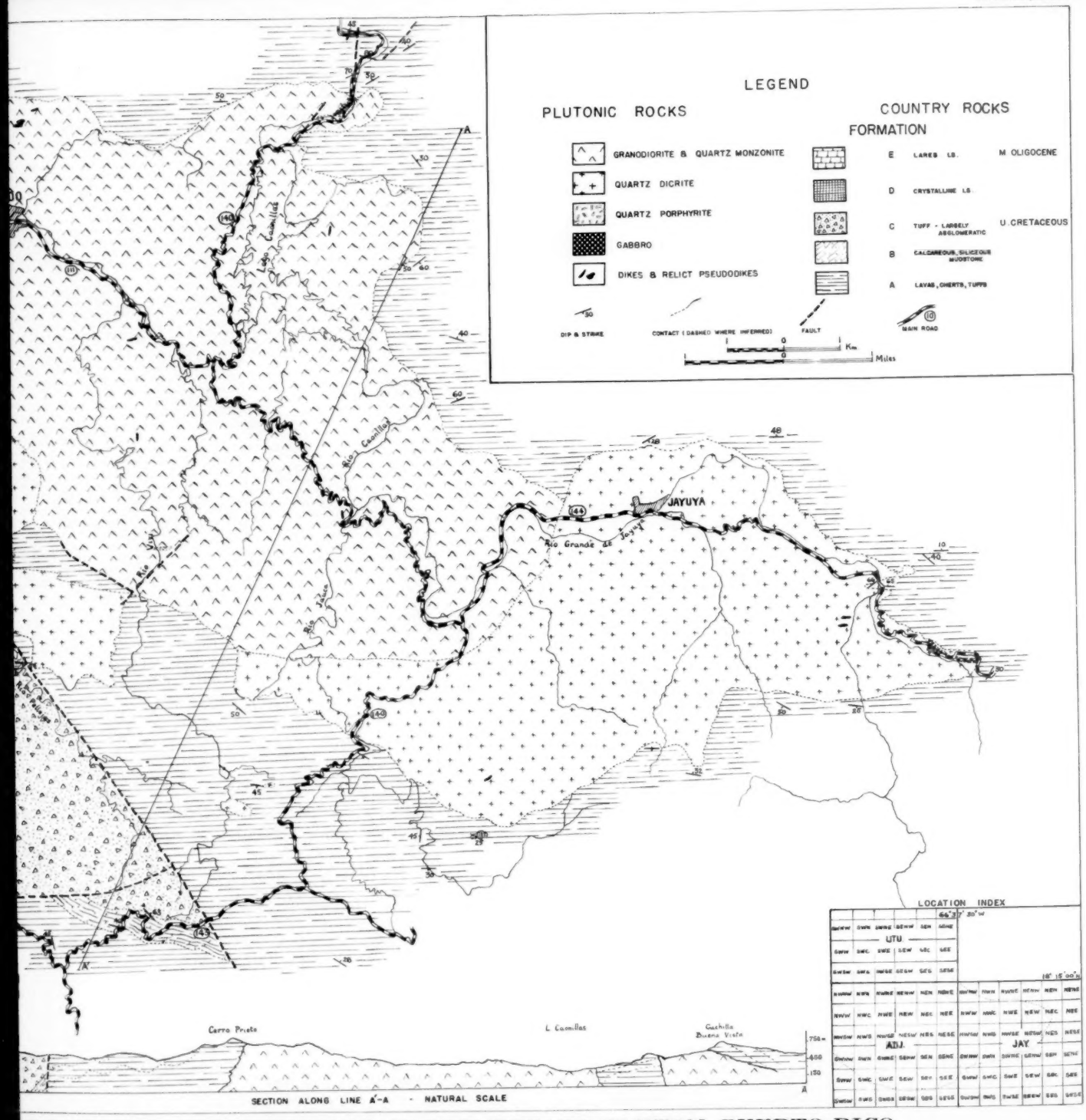

IC MAP AND CROSS-SECTION OF THE UTUADO PLUTON, PUERTO RICO

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UTUADO PLUTON, PUERTO RICO

BY JOHN D. WEAVER

ABSTRACT

The Utuado pluton is emplaced among Upper Cretaceous volcanic and sedimentary rocks in north-central Puerto Rico. It consists mainly of granodiorite and quartz diorite; minor amounts of quartz porphyrite and gabbro occur marginally. Present evidence favors the conclusion that these rocks were formed by transformation of pre-existing rocks, without, in the main, passing through a liquid state.

The age of the pluton is not yet accurately fixed but is probably late Cretaceous, although it may be early Eocene. The pluton probably was formed at not more than 3 miles depth and probably very much less.

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INTRODUCTION

This is a preliminary account of the Utuado pluton, situated in north-central Puerto Rico on the northern flank of the Cordillera Central. The pluton occupies approximately 130 square km and is surrounded for the most part by Cretaceous rocks. No detailed geological map of this region of the island has yet been published. The approximate outline of the pluton is shown on Meyerhoff's geological map of Puerto Rico (1933) and in maps by McGuiness (1948) and Mitchell (1954). The soils map of Puerto Rico (Roberts 1942) indicates the area underlain by the pluton. Kaye (1957) made some invaluable observations in the area, chiefly underground in the tunnels associated with the Caonillas reservoir.

The pluton consists of rocks ranging from granodiorite to gabbro in composition and seems to be an average-sized representative of the more acidic rock bodies of the Antillean region; this, coupled with its reasonable accessibility and exposure, renders it an appropriate starting point for a study of the dioritic rocks of the West Indies.

Rutten (1939) and Mitchell (1953) summarized the age problems involved. Only a few of the dioritic bodies have been studied or mapped on anything more than a reconnaissance basis.

The Antillean island arc is a region of recent and continuing tectonic activity, so that the petrology and structural and age relationships of the acidic igneous rocks assume some importance. Ewing (1953), for example, has suggested that these rocks have been formed by granitization of sediments which accumulated in an oceanic trough.

Acknowledgments

The original reconnaissance of the area was carried out as a minor aspect of a project of Prof. Walter Bucher of Columbia University and was supported by National Science Foundation Project No. G 442. The author is grateful for this original support and also to Professor Bucher for his continuing interest and encouragement. Work was continued with support from The Geological Society of America, which is also gratefully acknowledged.

Considerable assistance has been received from conversations with geologists now working in Puerto Rico, especially Prof. H. H. Hess and Drs. T. Slodowski and P. Mattson. Mr. E. A. Pessagno, Jr. examined the rather poor Foraminifera, and Dr. Raymond Smith and R. Gillou, J. Owen, R. Tichsler, and Watson Monroe of the U. S. Geological Survey have discussed field observations with the author. Mr. Mort Turner, Chief Geologist of the Economic Development Administration of the Commonwealth of Puerto Rico, has also been helpful.

Dr. C. A. Kaye and H. A. Meyerhoff read the manuscript and offered valuable criticism. They are not, of course, in any way responsible for any views expressed here. Mr. David Othmer made the photomicrographs.

Method of Locating Points

The area studied falls within three sheets of the 1:30,000 series of topographic maps—the Utuado, Adjuntas, and Jayuya quadrangles. These sheets are divided into four areas covered by appropriate 1:10,000 sheets which are named, *e.g.*, Adjuntas NE, NW, SE, and SW. For convenience the 1:10,000 sheets have been divided into nine squares which have been named NW, N, NE, W, C, E, SW, S, and SE; hence any square can be referred to by quoting the name of the major quadrangle and three or four letters to indicate the squares. Thus, Adjuntas NWNE will refer to the northeast square of the northwest 1:10,000 sheet of the Adjuntas quadrangle. Locations quoted thus will probably suffice for ordinary purposes; a key to the squares is given in the lower right corner of the geological map (Pl. 1).

For more accurate location of points, the appropriate 1:30,000 quadrangle must be consulted. This is divided into quarters; locations will also be quoted in tenths of an inch north from the southern boundary of these quadrants or south from the northern boundary, and east or west from the western and eastern boundaries respectively. Thus, ADJ.NE63S24E will designate a point on the Adjuntas quadrangle, in the northeast quadrant, 6.3 inches from the northern boundary of the quadrant and 2.4 inches from the western boundary. This method is similar to that employed by Weaver (1957).

Geography and Geomorphology

Figure 1 indicates the general location of the Utuado pluton. The pluton is situated on the northern edge of the Cordillera Central, a region of strong relief and extremely steep slopes (up to 50-degree slopes have been observed). In general the area underlain by the pluton is less rugged than the rest of the Cordillera; the topography is noticeably more rounded and of lower elevation, although most slopes are steep. The altitude of the hills in the central part of the pluton averages about 600 m; the surrounding part of the Cordillera averages 800–900 m and is 1100 m in high points. The topographic change does not follow the outline of the pluton in detail, however; many contacts occur part of the way up the steeper slopes which rise to the upper levels of the Cordillera without causing any noticeable feature. The area to the north, underlain by Tertiary limestone, exhibits well-developed karst topography and presents a pronounced south-facing escarpment.

The pluton is for the most part covered with sugar plantations and tobacco, and the surrounding areas are devoted mainly to coffee and a few citrus fruit plantations. Smaller areas, particularly toward the southeast, are under pasture.

The drainage is north to the Atlantic. There are several relatively large streams (the Río Grande de Arecibo is perhaps the most notable) and innumerable small tributaries. Most of the streams are actively downcutting. Waterfalls and rapids are common. Well-marked terraces and alluvial gravels at varying elevations above

the present watercourses have been observed but not studied. The Río Grande de Arecibo has a fairly wide flood plain in the vicinity of Utuado but narrows again into a steep-sided gorge to the north, where it passes through the escarpment of the Tertiary limestone. The Río

FIGURE 1.—INDEX MAP OF PUERTO RICO

Grande de Jayuya has a wide flood plain in the neighborhood of Jayuya, but the valley narrows again to the northeast.

There are three major towns in the area, Adjuntas, Utuado, and Jayuya. There are several large haciendas and one big sugar central, Central Pellejas, between Utuado and Adjuntas.

Lago de Dos Bocas (on the Río Grande de Arecibo north of Utuado) and Lago Caonillas (on the Río Caonillas)—both artificial reservoirs—are part of the hydroelectric scheme of the Puerto Rico Water Resources Authority.

In the northwestern part of the area a notable erosion surface occurs at an altitude of approximately 480 m. It is particularly well developed in the Barrio Caguana, where it forms a distinct plateaulike area. This is the type area of the Caguana peneplain of Meyerhoff (1927). This altitude is favored by many summits in the vicinity; the surface, which in the Caguana area is developed on strongly folded Upper Cretaceous rocks and also on diorites continues northward, beveling the top of the northern coastal plain (middle Oligocene) limestone.

At higher elevations, benches, accordant summit levels, and sharp changes of slope are common, and these features are under study. The upper part of the Cordillera, discounting the few peaks with the highest elevations, maintains a fairly consistent level at an average altitude of about 750–800 m and forms a noticeable skyline feature. This presumably corresponds to the St. John peneplain of Meyerhoff (1927).

On the whole, the region is surprisingly accessible. Most parts are approachable to within about half a mile by reasonably good roads. In most areas networks of footpaths and tracks enable one to traverse the country, although some parts are so steep and thickly covered with vegetation as to be impassable.

COUNTRY ROCKS

Description of Formations

The pluton is emplaced among rocks which have been referred to as the Older Series of Puerto Rico (Berkey, 1919), and which Meyerhoff (1933) considers to be entirely Upper Cretaceous. These rocks are mainly of volcanic origin, although a considerable volume of volcanically derived sediments and some limestones are present; the latter are more common in neighboring areas.

The rocks of the Older Series are folded, faulted, and mildly metamorphosed and in the north are unconformably overlain by the almost horizontal Lares limestone of the middle Oligocene. The plutonic rocks crop out near the latter but are not seen in direct contact with it.

The older group of rocks can be divided into four mappable formations. Since the U. S. Geological Survey is engaged in systematically mapping the entire island, the formations mapped here have not been named, as this would probably cause confusion later. Consequently formations are here designated by letters and can be named after the stratigraphic sequence of the island as a whole has been better established.

Formation *A* consists typically of a thick sequence of interbedded lavas and bedded, silicified siltstones and mudstones, most of which are dark gray or black. Some tuffs and agglomeratic rocks also are found. Formation *A* is the most extensively developed formation in the neighborhood of the pluton and is best exposed along the eastern boundary, particularly east and northeast of Lago Caonillas and Cerro Morales. Exposures on the western and northern sides of the pluton are scattered and in most places are badly weathered. Most exposures to the south are also poor, although fresh outcrops are found in a few places.

The two characteristic lithologies of this formation can be recognized even in most of the poorly exposed areas. The lavas, predominantly andesitic, commonly contain large porphyritic crystals of hornblende, some of which are as large as 2 cm. Porphyritic feldspar commonly about 2 or 3 mm in length is also present in most examples. Conspicuous feldspars are common in many of the rocks, in tuffs as well as in lavas, and are not therefore diagnostic.

Hornblende andesites approaching basaltic composition seem to be quantitatively the most

important among the lavas, but other types are also found, notably pyroxene basalt. The latter in most cases contains euhedral pyroxenes 1–2 mm in length. Some samples of the lavas are deficient in ferromagnesian minerals, at any rate as phenocrysts. However, it is impossible to be categorical as to the relative proportions of various rock types.

These rocks weather yellowish brown. Many are speckled with altered feldspars, and the ferromagnesian minerals remain visible as rusty patches even after extensive weathering.

The silicified mudstones and siltstones are conspicuously banded. They are hard, fine-grained, siliceous rocks, bedded in layers about 4–5 cm thick, and are in most cases a pale whitish green. The banding which parallels the bedding is due to color variation from pale green or white to dark green. The best exposure of these rocks is along Route 10 about 1¾ miles north of Utuado. (UTU.SWNE). Good exposures are found also on the flanks of Cerro Morales (JAY.NWNW). A fairly long section is also exposed along Route 140 northeast of Lago Caonillas (UTU.SEE,SENE). Here the rocks are dark gray or black, apparently because of mild metamorphism.

Silicified mudstones and siltstones are also found in smaller outcrops east and south of the pluton. In this general area in particular, boulders of this lithology are common as float and in stream gravels, indicating that the rock is probably much more extensive than the limited exposures suggest.

In fresh samples, some small patches react with dilute HCl; this probably indicates the remains of Foraminifera, for in thin section tiny circular indeterminate calcite patches are found. In most cases leaching has removed the calcite.

The silicified mudstones and siltstones appear to resist weathering somewhat more than the lavas; even so they are extensively weathered, with leaching of silica and production of a yellowish-brown friable material. This material, however, retains traces of the thin bedding until weathering is highly advanced.

While it is not possible to state categorically whether silicification is primary or secondary, it probably is due to low-grade metamorphism and recrystallization of original mudstones or siltstones. In some outcrops—*e.g.*, NE of Lago Caonillas—some layers have coarser grains. Feldspar grains and few rock fragments are also found, suggesting that the rocks were either tuffaceous or graywackes.

In addition to these two major lithological types, two other types are found in scattered outcrops. Their inclusion in formation *A* seems justified on the basis of their geographical relationship.

The first of these is a conspicuous rock consisting of alternating layers of purplish-red tuffaceous siltstone and harder and thinner layers of greenish, siliceous tuff. The purple layers are calcareous in spots, and in one place definite organic fragments were found; these appeared to be foraminiferal remains. This lithology is well exposed in the valley on the north side of Mt. Hormiga (ADJ.NWW) but elsewhere is found only in isolated patches, one in the southeast corner of ADJ.SEN (ADJ.SE-24SIW) and another in a quarry in ADJ.SWE. (ADJ.SW40N8W).

The rocks exposed in the Mt. Hormiga area dip under formation *B*; therefore it is inferred that this lithology occurs at the top of formation *A*.

The other lithology is also reddish purple but consists of interbedded siltstone, sandstone, and conglomerate. In the conglomerate most of the fragments are angular and consist predominantly of andesite; many of them display small but conspicuous purplish or gray feldspar phenocrysts. One or two chert pebbles are also found. In one or two places, the conglomerate, which occurs in irregular lenses, rests on a scoured surface of sand, which shows poorly developed current-bedding in some places. This lithology can best be seen in the bed of the Caonillas River, just above the Caonillas No. 2 hydroelectric plant at UTU.SE22S22W. A poor exposure is found at JAY.SW29S10E.

Thickness estimates in terranes of this kind are generally little more than controlled guesses. The section northeast of Lago Caonillas displays approximately 1000 m of bedded silicified mudstones and siltstones with a few interbedded lavas. At Cerro Morales, fairly regularly interbedded lavas and siltstones, the former averaging about 8–10 m thick, total approximately 1100 m. In neither case is there any evidence of strong faulting; dips are fairly uniform at about 45°–60° north. The differing proportions of lava suggest that there is little if any overlap between the two sections, and consequently the thickness of formation *A* is estimated as not less than 1500 m or more than 2500 m. About 2000 m seems a reasonable estimate.

Overlying formation *A* is a relatively thin but distinctive series of calcareous mudstones—formation *B*. The mudstones are distinctly stratified in beds 2–6 inches thick and characteristically are light buff or light gray. They are siliceous in places. The bedding is evident even

after weathering, although the rock becomes friable and the calcium carbonate is leached out. Nevertheless the mudstones do represent a recognizable and mappable horizon and have hence been treated as a separate formation, even though they may merely mark the top of formation *A*, differing only in the higher proportion of calcium carbonate present.

Weathering and scattering of outcrops make it difficult to determine thickness, but the thickness is estimated at approximately 100 m.

E. A. Pessagno, Jr. collected some poorly preserved Foraminifera from a roadcut on Route 143 (ADJ.SE41N7E) and identified them as *Globotruncana calcarata* and *G. lapparenti* (?), probably Campanian.

Formation *B* is also well exposed in a roadcut on Route 10 at ADJ.NW55S29E). (Southwest corner of ADJ.NWC).

Formation *C* consists of a conspicuous breccia. In fresh cuts, it has a bluish-gray, moderately coarse, tuffaceous matrix, in which are scattered fragments almost exclusively of purple or gray andesite. The fragments range from about 2 cm to 30 cm or more and are in general angular or subangular, mostly with rounded edges. They are scattered through the matrix in such a manner that they appear to be suspended in it. The matrix is calcareous in spots. The proportion of larger fragments varies, and in some places they are absent.

Formation *C* is widely distributed, and owing to its resistance, it forms a ridge extending from west of Mt. Hormiga, whose crest it underlies, eastward along the Cuchilla Juan González.

It weathers to a light brownish-red clay with large, almost fresh, rounded residuals scattered through it. Landslides are common where roads have been cut through it.

The rock is easily recognized in the field in most places, but in thin sections it seems to be obscured by alteration. The fragments are not easily distinguished from the matrix, and therefore it would seem that the matrix and fragments are composed of the same material, that is, volcanically derived andesitic material. The calcite in the matrix probably indicates the former existence of foraminifers, although none have been found. Turner (1955, p. 6) reports Foraminifera and other organic material in a similar rock in the San Sebastián area. This suggests that the rock is not an agglomerate in the usual sense of the word, and so the more general term breccia is used here. The rock was deposited under marine conditions; the massive form, general lack of bedding, and the curious poorly sorted scattering of the larger slightly rounded fragments seems to indicate that it was formed by submarine slumping. Turner (1955, p. 6) has suggested an origin as a submarine flow or lahar. However, the fragmental nature suggests that it was more probably formed by slumping into deeper parts of the sea from shallower areas where it originally accumulated.

Toward the bottom, large fragments are absent; in surface outcrops some indistinct bedding is found. Kaye (1957, Pl. 4) indicates more distinct stratification in the tunnel section which just penetrates the lower part of the formation.

Just northeast of Adjuntas (ADJ.SW22S-41W), well-bedded volcanic wacke is exposed. It seems to be interbedded with the breccia and is believed to be near the top of the formation. It is lithologically identical with the bedded wackes of the Yauco formation (Slodowski, 1956, p. 63; Mattson, 1957, p. 80). It is medium gray with beds 5–10 cm thick and in some places exhibits graded bedding. In general the grain size varies from silt to sand, with poorly sorted angular to subangular grains. It contains considerable plagioclase feldspar, quartz, calcite, and clay-sized material. The grading occurs within a thickness of about 2 cm and in most cases takes the form of a 1–2 mm band of coarse particles (about 1 mm in diameter) with much interstitial clay, grading upward into silt size.

Possibly this is part of the overlying formation, but in view of the very small area of outcrop observed, it is included in formation *C*.

North and northwest of Adjuntas, formation *C* crops out extensively. The width of outcrop is a little over 3 km; dips over the whole outcrop area, where observable, are fairly uniform, averaging 40°–50°. There is no reason to suspect any large-scale repetition by faulting, and hence a thickness of 2000–2500 m is indicated. The rocks below show considerable small-scale folding, and this may also affect the breccia; owing to its massive form, this is not in evidence. A considerable thickness (conservatively estimated at 1500–2000 m) is present.

Formation *D* crops out in one very small area in Jayuya SWNW. It is clearly unconformable to the underlying rocks of formation *A*. As there is no reason for suspecting any unconformity between formations *A*, *B*, and *C*, this limestone is presumably younger than and unconformable upon all the previously described formations. It is a distinctly bedded, whitish-gray, coarsely crystalline limestone with thin (5–6 mm) sandy layers spaced about 2–3

cm apart. These latter are somewhat crinkly or plicated irregularly in small folds about 1 cm in amplitude. No recognizable fossils were found, but there are considerable organic fragments.

Formation *E*, the Lares limestone, is a white to buff hard limestone of middle Oligocene age (Zapp *et al.*, 1948), which lies almost horizontally and with strong angular unconformity above all the previously described formations. It has not been studied in connection with the present work. It forms a strong escarpment in the northern part of the area and gives rise to rugged karst topography.

Stratigraphy and Correlations

Since other systematic mapping has not yet nearly approached the area under consideration, and since this area is not ideal for setting up a stratigraphic section, any correlations indicated here are naturally tentative.

To the west, Hubbard (1923) described a large region extending from the west coast to the vicinity of Lares, about 15 km east of Utuado. This work was of a reconnaissance nature, but he was able to distinguish three formations below the Tertiary coastal plain limestones. These he named in order of their succession: the Río Culebrinas series, the Río Blanco series, and the Río Yauco series. However, recent observations have thrown some doubt on this relationship. (Mattson, 1957 p. 130).

Turner (1955) has more recently mapped in detail an area in the neighborhood of San Sebastián, west of Lares, where he recognized the two latter series of Hubbard. However, the same doubts apply to the stratigraphic correlations adopted by him.

South of Adjuntas, Slodowski (1956) has mapped and described a large area centering on Yauco and extending north to within 5–6 km of the latitude of Adjuntas. Unfortunately, most of his work was done in the southern part of his area, and the northern part was covered only by reconnaissance. In the north he recognizes the Río Yauco and Río Blanco, which he more properly renames formations.

West of the Yauco area, Mattson (1957) has mapped the Mayagüez area, which extends southward from Mayagüez (situated centrally on the west coast) to the south coast and eastward to join up with Slodowski's area. He has renamed the Río Yauco formation the Yauco mudstone and also recognizes the Río Blanco formation in the northern part of his area, but leaves the age relationship of the two in doubt.

Slodowski (1956, p. 82) considers the Río Blanco and Río Yauco to be laterally interfingering time-equivalents.

Obviously there is no good standard section with which to attempt the correlation of the rocks in the Utuado area. However, several observations may be offered as a basis for possible correlation.

The calcareous, siliceous mudstones of formation *B* strongly resemble the topmost beds of the Río Culebrinas formation in the San Sebastián area (Turner, Personal communication). The bedded wacke, which is apparently at the top of the breccia unit, formation *C*, is lithologically identical with the typical rocks of the Yauco formation exposed near Rincón, in the northwestern part of the island, and also described by Slodowski (1956, p. 66) from the Yauco area.

The small extent of the outcrop of formation *D* makes it difficult to correlate, but the limestone does bear similarity to some limestone of the San Germán formation (Slodowski, 1956, p. 86); its unconformable relationship to the older rocks supports such a tentative correlation.

Both Mattson (p. 130) and Slodowski (p. 82) find interpretation of the stratigraphical relationship between the Río Blanco and the Yauco formations difficult. Slodowski concludes that they are partly, at least, laterally interfingering facies. In the present state of our knowledge, it seems reasonable to correlate the breccia formation *C* with the Río Blanco; the bedded wacke at the top may possibly represent the bottom of the Yauco. The scanty foraminiferal evidence from formation *B* is not inconsistent with this.

The following stratigraphical correlation is adopted tentatively until mapping is extended from areas where the stratigraphy can better be established:

E		Lares limestone	Middle Oligocene
D		? San Germán limestone	Upper Cretaceous
C	Breccia (wacké above)	Yauco formation Río Blanco formation	Upper Cretaceous
B	Calcareous siliceous mudstones	Río Culebrinas formation	Upper Cretaceous
A	Lavas and cherts		Upper Cretaceous

The variation in the apparent thicknesses supports Slodowski's and Mattson's conclusions as to the probable relationship between the Yauco and Río Blanco formations. It seems also to apply to the Río Culebrinas formation. That is, the three units in bulk are contemporaneous, but the locale of formation of the differing lithologies—breccias, tuffs, flows—shifted from time to time from one area to another, with the result that similar lithologies are not necessarily equivalent in time.

Pluton Phases

General

The pluton occupies an irregular area approximately 20 km long by 8 km wide, extending east-southeast from west of Utuado to east of Jayuya.

It varies in appearance and composition, and four main rock types have been recognized and mapped. These are: (1) granodiorite and quartz monzonite; (2) quartz diorite; (3) quartz porphyrite; and (4) gabbro.

The granodiorite and the quartz diorite are quantitatively by far the most important; the quartz porphyrite is a border phase. The gabbro also occurs in very small bodies at the edges of the pluton.

Several small bodies of dioritic rock are found outside the pluton. They seem to be lenticular and are regarded as offshoots of the main body.

These rocks retain their texture even after intensive weathering. Most of the plutonic rocks can be recognized, even though they are deeply weathered; in most cases it is not possible to distinguish the different phases, although in some cases they can be distinguished on a textural basis.

Few contacts are exposed. At the boundaries between the pluton and the country rocks, mixed zones of the plutonic rock and irregular residual masses of the country rocks are found. No sharp contacts between the phases of the pluton have yet been found. Wherever they come together there appears rather to be a gradual transition. The granodiorite and the quartz diorite are distinguished in the field by visible pink orthoclase in the former; generally speaking, in any of the contact zones it is difficult to decide exactly where to place the boundary. Figure 2a-e shows thin sections of typical examples of the various phases.

Granodiorite

Granodiorite forms the main central mass of the pluton. In fairly fresh exposures it is a massive, jointed, light-colored rock, with a medium to coarse granitoid texture. It is more or less equigranular and contains visible quartz, plagioclase, pink orthoclase, and hornblende. Considerable biotite occurs locally, but it is not widespread; altered material, in most cases chlorite, is common and may be derived from biotite. Quartz makes up 20–25 per cent of the granodiorite. The plagioclase has an average composition of An_{35}–An_{40}. Green hornblende occurs in stubby crystals.

The granodiorite varies considerably particularly in the amount of orthoclase, which diminishes toward the contact with the quartz diorite, so that the transition from one to the other is not sharp.

Table 1 shows that the granodiorite is fairly close to the average granodiorite of Daly (1933, p. 457), although the silica content near its borders is low for some definitions, and it becomes a quartz monzonite or, where the orthoclase content is low, a tonalite.

The rock is exposed in many outcrops in the vicinity of Lago Caonillas, particularly near the dam, and there are several good exposures along Route 140 between Utuado and Jayuya.

Quartz Diorite

Quartz diorite crops out marginally on the western, southwestern, and southeastern sides of the granodiorite.

Characteristically it is a massive, white-appearing, granitic-textured, medium-grained rock and contains conspicuous lathlike hornblendes.

This rock is distinguished in the field from the granodiorite by the absence of visible orthoclase and by the form of the hornblende, which in the granodiorite is more stubby. The quartz diorite varies in appearance. In some places it seems to be darker in color, with hornblendes less conspicuous. In thin section, nevertheless, there is no notable difference.

Essentially the quartz diorite consists of plagioclase feldspar of composition about An_{35}–An_{50}, quartz, and hornblende. Apatite and sphene are the chief accessory minerals besides the opaque minerals—mostly magnetite—and an occasional zircon. Biotite is found in some places.

In general the plagioclase is euhedral and displays polysynthetic twinning. Irregular zoning, in most cases normal or oscillatory normal, is fairly common, and former crystal boundaries are commonly outlined in sericite. Quartz is anhedral and averages about 25 per cent of the rock, although in some cases it is as low as 10

FIGURE 2.—THIN SECTIONS OF TYPICAL ROCKS OF THE PLUTON

a, granodiorite; b, quartz diorite; c, quartz porphyrite; d, gabbro; e, andesite from relict pseudodike. (In the figures Q-quartz, Pl-plagioclase, Or-orthoclase, Hb-hornblende, Py-pyroxene.)

per cent. Hornblende is pleochoric, medium green to pale straw yellow. The texture is hypidiomorphic granular with grain size averaging 1–2 mm, although some of the hornblende and plagioclase crystals attain lengths of 3–4 mm or more.

TABLE 1.—CHEMICAL ANALYSIS OF UTUADO PLUTONIC ROCKS
Analyst H. B. Wiik, Helsingfors, Finland

	No. 1	No. 2	No. 3	No. 4	No. 5	No. 6	No. 7	No. 8
SiO_2	46.21	63.28	64.81	59.10	61.30	64.94	66.70	62.70
TiO_2	1.07	0.47	0.41	0.43	0.35	0.31	0.45	0.57
Al_2O_3	18.92	16.61	15.40	18.45	17.27	16.71	16.15	17.67
Fe_2O_3	4.97	3.35	3.02	2.84	2.01	2.51	2.10	2.95
FeO	7.19	2.77	2.06	3.20	3.34	1.92	1.86	2.82
MnO	0.20	0.10	0.07	0.14	0.08	0.07	0.08	0.14
MgO	6.45	2.38	2.07	2.75	2.12	1.30	1.72	2.11
CaO	12.49	5.64	4.91	6.77	4.48	4.99	4.23	5.85
Na_2O	1.86	3.18	3.18	3.13	3.16	3.69	3.60	3.26
K_2O	0.33	1.28	2.20	0.97	1.48	1.31	2.49	1.53
P_2O_5	0.15	0.15	0.15	0.14	0.11	0.13	0.04	0.02
H_2O^+	0.50	0.92	0.61	1.25	2.57	0.98	0.26	0.40
H_2O^-	0.01	0.05	0.02	0.05	0.12	0.08	0.02	0.02
CO_2	0.00	0.00	0.49	0.00	1.21	0.84	0.00	0.00
	100.35	100.18	99.40	99.22	99.60	99.78	99.70	100.04
Norms								
Q	—	24.06	25.98	17.10	26.22	30.24	24.42	21.18
or	2.22	7.78	12.79	6.12	8.90	7.78	14.46	8.90
ab	15.72	26.72	26.72	26.72	26.72	31.44	30.39	27.77
an	41.98	26.97	20.29	32.80	13.90	18.63	20.57	29.19
C	—	—	0.41	0.10	5.41	2.35	—	—
di	15.48	0.22	—	—	—	—	0.46	—
hy	11.26	7.12	5.86	9.94	9.13	4.12	5.12	7.28
ol	3.95	—	—	—	—	—	—	—
mt	7.19	4.87	4.41	4.18	3.02	3.71	3.02	4.41
il	1.98	0.91	0.76	0.76	0.76	0.61	0.91	1.05
ap	0.34	0.34	0.34	0.34	0.34	0.34	0.06	0.03
cc	—	—	1.10	—	2.70	1.90	—	—
Symbol	(II)III5 4(5)(4)5	(I)II 4(3)4 4	(I)II 4 3 4	'II 4 "4 4"	"II (3)4 3 4	I(II)"4 3 4	I(II) 4 3 "4	(I)II 4 3(4) 4
	Auvergnose	Tonalose	Tonalose	Bandose	Tonalose	Yellow-stonose	Yellow-stonose	Tonalose

No. 1—gabbro, CAR. 52, Locality JAY. NE18N20E; No. 2—quartz monzonite, CAR. 58, Locality JAY. NW45S20E; No. 3—quartz monzonite, CAR. 59, Locality JAY. NW1W13N; No. 4—quartz diorite, CAR. 63, Locality ADJ. NE58S13E; No. 5—andesite (relict pseudodike), CAR. 65, Locality ADJ. NE44S24W; No. 6—quartz porphyrite, CAR. 64, Locality ADJ. NE41N19E; No. 7—granodiorite, CAR. 27, Locality ADJ. NE8W38N; No. 8—quartz diorite, CAR. 141, Locality JAY. NW43N16W.

Quartz Porphyrite

Quartz porphyrite forms a marginal phase of the pluton and is found in small areas interfingered with the quartz diorite, chiefly along the western margin, but it is also found in small patches in the southern part.

It is a medium gray, fine-grained rock with conspicuous, more or less circular spots of white plagioclase and some quartz. However, in some exposures the porphyritic nature is not evident, and the rock then looks like the darker variety of the quartz diorite. Its contact with the quartz diorite is transitional.

It consists essentially of plagioclase feldspar, quartz, hornblende, and accessory opaque minerals. It has a fine- to medium-grained groundmass with a granoblastic texture, and enclosed in this, large crystals of plagioclase and quartz. In specimens taken across contacts with the quartz diorite, the grain size of the groundmass increases until the size of the larger crystals is attained; at this point, the quartz porphyrite becomes indistinguishable from the quartz diorite.

The feldspars show all stages of zoning from strong oscillatory normal to a very faint residual zoning. Twinning is also found in all stages of development. Average composition is ap-

proximately An_{48}. Quartz tends to occur in rounded grains, of pool-like form; in places re-entrants enclose part of the groundmass. Some euhedral crystals also are found.

Gabbro

Gabbro is quantitatively the least developed of the coarser-grained rocks of the pluton. It is distinctly marginal, occurring in small bodies at one or two points at the edges of the pluton.

In gross appearance it is not noticeably different from the diorites, but freshly broken surfaces are darker gray and show conspicuous and in some cases large (up to 0.5 cm) pyroxene phenocrysts. The gabbro is fairly coarse-grained, with panidiomorphic texture. It consists essentially of plagioclase (about An_{61}) and augite. The latter commonly has coronas of hornblende and some biotite. Small amounts of hypersthene are also present.

Minor Bodies

Minor rock bodies within the pluton are of three types: (1) parallel-sided dikes; (2) irregularly shaped dikelike bodies, interpreted here as relict pseudodikes; and (3) thin aplite veins.

Most definite parallel-sided dikes are not well exposed. They are probably more numerous than surface exposures would suggest. In UTU.SWC one of dark-gray andesite cuts the granodiorite. It is 5 m wide and strikes northwest. At UTU.SEC a small basaltic dike about 45 cm wide strikes about north-northwest, and at UTU.SWSW is a small andesitic dike about 1.25 m wide, which strikes northeast.

Numerous parallel-sided bodies which have similar lithology to the bodies of type 2 are associated with some of the phenomena mentioned below. Most of them are badly weathered and are difficult to categorize. For example, in ADJ.NEE there are deeply weathered but easily visible porphyritic dikes, of the order of 1 m thick and parallel-sided, which are of apparently identical lithology to the rocks of type 2; near these are rounded xenoliths, also of the same lithology, enclosed in the granodiorite.

Dikelike bodies are common and uniform in appearance. They are composed of a rock which has the appearance of porphyritic andesite; it is fine-grained, dark gray to black, and speckled with small feldspar phenocrysts.

These bodies are irregular, pinching in and swelling, and interfingering with the surrounding diorite so that it is impossible to say which is the intrusive (Fig. 3). In places small ovoid diorite masses are enclosed in these rocks; in others small subcircular pieces of the porphyritic andesite are found in the diorite. Feldspar porphyroblasts transgress the contacts. These observations lead to the conclusion that the bodies are relict pseudodikes (Goodspeed, 1955).

These relict pseudodikes are common in all parts of the pluton, especially near the boundaries. They are conspicuous even when deeply weathered, and they weather more rapidly than the surrounding rocks, so that fresh samples are rare.

Aplite veins are common in some parts of the pluton; slightly more are present in the granodiorite than in the quartz diorite. In most cases they are only 4–5 cm thick but have considerable lateral extension; the full extent cannot be estimated, since they run the whole width of exposures. They maintain a uniform thickness and straight walls.

A thin section from one taken from near the contact zone at UTU.SEC. (UTU.SE43S28W) shows that it consists of a fine granular mass of quartz, plagioclase, and orthoclase, with extensive, well-developed myrmekitic intergrowths of quartz and plagioclase. The latter is approximately an oligoclase. In hand specimen most are of a fine sugary texture and white, commonly with a slight pink tinge.

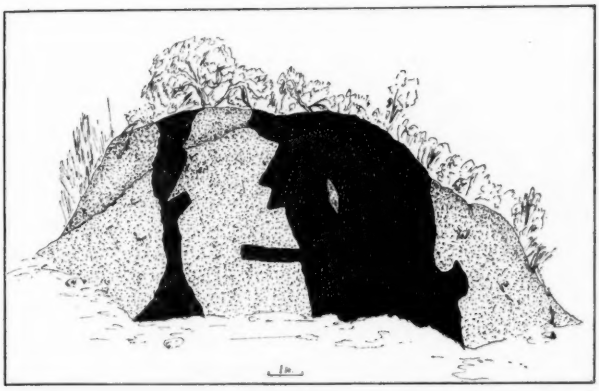
Xenoliths

Xenoliths are common in both the granodiorite and the quartz diorite phases but appear to be more common in the former.

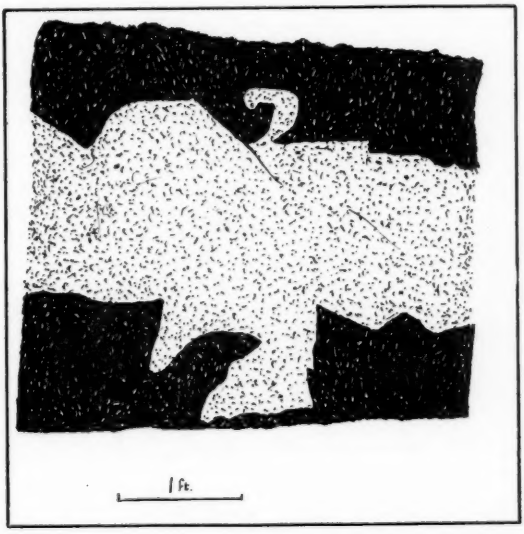
They consist for the most part of small rounded or subrounded fragments, 4–5 cm across (some are up to 30 cm), of fine-grained, dark gray or black material. Very commonly they are peppered with porphyroblastic plagioclase crystals, which, in many cases, extend across the contact. These appear to be lithologically identical with the relict pseudodikes described above.

Some xenoliths which do not have the conspicuous feldspar porphyroblasts are found on thin-section examination to have the same mineralogical composition as the enclosing diorite and differ only in having a much finer grain size, with perhaps a slight dominance of hornblende.

One xenolith at JAY.NWS (JAY.NW7-N35W) a little over half a kilometer from the contact is a fragment of only slightly altered, banded silicified mudstone, characteristic of formation *A*. It has a dark, fine-grained selvage,



a



b

FIGURE 3.—SKETCH OF RELICT PSEUDODIKE AT UTU. SW34S27W AND DETAIL OF ANOTHER

which microscopically is seen to consist of a fine granular mass of pyroxene, quartz, and plagioclase—identical with the hornfels to be described. The transition at the contact to diorite is not sharp but the grain size does increase within a short distance (Pl. 2, fig. 5). Pyroxene occurs in the diorite near the contact but is progressively replaced by hornblende farther into the diorite. Large quartz and feldspar crystals enclose granoblastic-textured minerals of the hornfels. Biotite is also present near the contact. Preservation of banding in the hornfels indicates that the rock was never liquid during recrystallization.

These observations suggest that the xenoliths are residuals of country rock altered through a hornfels stage into a material identical with the diorite but with finer grain size.

In the roadcuts northeast of the Caonillas dam, many large xenoliths are found; all are dark and fine-grained and contain more or less porphyroblastic feldspar; all stages can be found, from those with few porphyroblasts to ones with such thickly disseminated feldspars that

only ghostlike residuals of the original fragments can be seen; in the latter the rock is almost indistinguishable from the granodiorite. Some of the less altered examples have sharply defined basic rims about 3–4 mm wide.

Such xenoliths would probably be explained as stoped fragments of the roof in various stages of solution in the magma, but this process is difficult to understand, first because the transformation seems to begin within the xenoliths and, second, if the magma were granodiorite in composition, the magma with dissolved fragments would have to have a different composition, whereas the ghost residuals are identical with the normal granodiorite.

Metamorphism of the Area

The rocks of the Older Series in Puerto Rico commonly exhibit low-grade metamorphism; large proportions of the rocks could be referred to as greenstones; the commonest phenomenon is chloritization.

In the vicinity of the pluton most of the rocks are dark.

The cherty rocks of formation *A* in a narrow band in the vicinity of the contact become a quartz pyroxene hornfels, a hard black rock, showing a faint residual banding on weathered surfaces. In thin section this rock is found to consist of a mosaic of quartz and pyroxene with some plagioclase. The banding is due to layers in which pyroxene is concentrated. (Pl. 2, fig. 4).

The more calcareous horizon of formation *B* is slightly marmorized near the diorite and assumes a light gray or buff color. It, too, is highly siliceous.

It has not been possible to ascertain exactly what happens to the andesitic and basaltic lavas. Thin sections of several dark, fine-grained rocks from near the contacts have the textures and general appearance of pyroxene or hornblende andesites with greater or lesser patches of granular quartz and some pyroxene or actinolite with a granoblastic or in some cases cataclastic texture. It is believed, therefore, that the lavas tend toward the same quartz pyroxene hornfels that results from metamorphism of the banded cherts.

Minor alteration is prevalent. In several localities rock zones have been reduced to a white or yellowish fine powdery material which has not yet been investigated. These zones seem to be along faults—as for example along the road to barrio Viví (ADJ.NESW) and in the road leading to Caonillas No. 1 hydroelectric plant (UTU.SENE).

Epidote veins are common in many parts of the area, particularly near contacts. Pyrite and chalcopyrite are also common but are not present in high concentrations. Small patches of azurite and malachite are found in scattered localities.

These effects seem to be mainly subsequent to the emplacement of the pluton, although some may be late-stage effects.

Structure of the Area

The region has been intensely deformed. The immediate neighborhood of the pluton does not seem to have as complex a structure as the Yauco area to the south or the Mayaguez area to the west; nevertheless it has undergone extensive folding and faulting.

In nearly all exposures the country rocks dip steeply, but on the whole, dips are fairly uniform over large areas, and the folds are therefore large scale. Small-scale folding or crumpling is observed in one or two places in the southern part of the area, but this is localized and probably associated with movement on faults. Kaye (1957, Pl. 4) indicates small-scale folding in the Adjuntas-Pellejas tunnel; the folding is not evident at the surface in all places. Clearly, deformation is more widespread than the surface exposures suggest.

The pluton is emplaced along the axis of a large anticline with a west-northwest axis, plunging southeast; the country rocks dip off the pluton to the north, to the south, and around the eastern end. This anticline is approximately 15 km across and is partly cut off in the south by an east-trending fault.

Faulting is extensive, although faults are not easy to trace in the field. Those shown on the map are either drawn from aerial photographs or inferred from displacement of outcrops or both. Fault zones and minor displacements can be seen in roadcuts, but many cannot be traced.

Slickensides are common; they are not associated with obvious faulting in every case and are probably produced by minor adjustment of the rocks to broader displacements. Most of the slickensides are horizontal.

The displacement on most of the larger faults is believed to be transcurrent or strike-slip.

Most faults trend east. The large fault south of Adjuntas is part of an extensive system of east-west faults which Slodowski (1956) has referred to as the Cordillera fault. It is a sinistral strike-slip fault.

Sets of smaller faults with northwest and northeast strikes are also prominent, and dikes, where found, also follow these directions.

The pluton itself is lenticular and trends generally northwest-southeast, as do the older rocks.

The attitude of the contacts is not clear, but the northern contact in the vicinity of the Caonillas dam appears to be inclined gently northward under steeply dipping country rocks (Kaye, 1957, p. 106). At most places on the surface where the contact is exposed, it is irregular and consists of a wide area of mixed rocks with large, irregularly shaped masses of dark, fine-grained hornfels mixed with diorite, all mutually interfingering and crosscutting. Such mixed zones can best be seen at the bridge across the Río Grande de Arecibo about 2 km north of Utuado and also at a small footbridge across the same river at ADJ.NWN (ADJ.NW20S35E). Similar irregular contacts can be seen in roadcuts northeast of Caonillas dam.

The major form of the pluton itself is also irregular; several large re-entrants of country rock extend into it, notably in the Central Pellejas (ADJ.NWE,NEW), near Coabey (JAY.NESW), and northeast of Lago Caonillas (UTU.SEC,SEE).

Several small lenticular or dikelike diorite bodies south of the pluton seem to be roughly aligned with the trend of the main pluton.

There are two well-defined unconformities in the area, one below the middle Oligocene Lares limestone, and the other below the small area of limestone of formation *D*.

The former is obvious and well known. The very fossiliferous white or buff, unaltered limestone rests almost horizontally, or with a small northward dip, on the steeply dipping formation *A*. The plutonic rocks are not found in direct contact at the unconformity, but the granodiorite is exposed in barrio Caguana (UTU.SWNW) within not more than 0.4 km of the Lares limestone. It may be much closer, for it is difficult to determine the exact boundary of the Lares limestone at this place, as the limestone has been spread by wash south of the escarpment. In any event the limestone is apparently unaffected by metamorphism, and it seems fairly definite that the pluton does not cut up into it.

The second unconformity is probably the same as that recognized by Slodowski (1956, Fig. 2) below the San Germán formation, although without further evidence for the correlation of the limestone, this is little more than conjecture.

The limestone of formation *D* dips about 25° southwest and lies on tuffs and agglomerates of formation *A*, which here are vertical and strike east.

Unfortunately within the area so far investigated the unconformity cannot be related to the emplacement of the pluton.

Interrelationship Between Phases of the Pluton

Lack of good exposures at critical points renders it difficult to be categorical about the contact relationships between the major phases of the diorite; however no sharp contacts have been found between the granodiorite and the quartz diorite. Where these two rock types approach one another in the field, it has been found that the boundary is difficult to place, and hence it is believed that the contact is transitional. This is borne out by the thin sections and chemical analyses (Table 1).

The conclusion that the contact is transitional is reinforced by the fact that the quartz porphyrite-quartz diorite contact is definitely transitional.

A series of thin sections taken across the contact between these two types at ADJ.NWW (ADJ.NW47E40S) shows the transition from the quartz porphyrite with granoblastic-textured groundmass (resembling hornfels) to typical quartz diorite by increase in the number of the large porphyroblastic feldspars and quartz, and increase in grain size of the groundmass.

Both the quartz porphyrite and the relict pseudodikes then, display transitional contacts with the quartz diorite and the granodiorite. The relationship of the small gabbro bodies is not clearly seen, but north of the dam at Lago Caonillas the gabbro appears to cut the quartz diorite with a sharp contact.

Origin of the Pluton

Several lines of evidence indicate that the diorites were formed by transformation of preexisting rocks, and there is some evidence to suggest that the transformation took place largely without the rocks passing through a liquid phase.

These are summarized as follows:

(1) Contact relationships. In addition to the features already described it should be mentioned that in several places near the contacts, the country rocks contain fairly conspicuous feldspars, and, although no good examples have been found to establish the fact, some at least appear to be porphyroblastic.

(2) Relict pseudodikes and xenoliths. These have been shown to be relics of country rock in process of transformation.

(3) General structural relationship. Two aspects of the structural nature of the pluton bear on the problem. The general concordance of the form of the pluton with the structure resembles bodies which have been interpreted as metasomatic rather than those which are definitely intrusive.

Also, the fact that the country rocks dip north, south, and east off the pluton can be explained in several ways. It might be argued that this indicates a forceful uprising of magma accompanied by the pushing aside of the country rocks. However, to the west this relationship does not hold good, for there the rocks strike into the pluton. More probably the diorites were emplaced in an already existing anticline. Even if it is contended that the structure shows a forceful pushing aside of the country rocks, this could as well be brought about by volume changes as by magmatic intrusion.

(4) One or two details of texture and mineralogy also support the view that the rocks were formed without passing through a liquid stage. The existence of porphyroblasts near contacts enclosing parts of the groundmass of hornfelsed country rocks has already been described; many feldspars in the diorites contain cores which show ghostlike relics of the original granoblastic groundmass. (Pl. 2, figs. 2, 3).

In tracing the transition across the contact between quartz porphyrite and quartz diorite, along with the increase in the number of feldspar porphyroblasts, there is also a progressive change in the zone-twin relationship. In the typical quartz porphyrite, zoned feldspars are the rule with, in a few cases, some fine, superimposed, irregularly developed twinning. As the over-all textural change progresses, the degree of twinning increases; in the typical quartz diorite twinning is the rule and zoning is commonly only faintly perceptible. This reconstitution of the feldspars seems, then, to have taken place in the solid state.

Some feldspars in the quartz porphyrite show from their form that they were fragments and not complete crystals, and the zoning parallels the fractured form. As the rock does not show any other signs of cataclasis, it would seem that it was derived from a rock containing feldspar fragments, *e.g.*, either a coarse tuff or breccia.

The details of the history of the feldspars and the ferromagnesian minerals are clearly significant to the problems of the genesis of these rocks; further studies are contemplated when the necessary equipment becomes available.

(5) Chemical analyses. Table 1 shows that the eight analyses of the typical rocks of the pluton all fall fairly close together, with the exception of the gabbro; the group as a whole is on the borderline of granodiorite and diorite. The Harker diagram (Fig. 4) demonstrates that there is no suggestion of straight-line variation, and therefore the rocks do not represent a differentiation series. It is suggested that they represent rather differences in the parental material.

Slodowski (1956, p. 110*a*) and Mattson (1957, p. 16, 23) have published analyses of 12 rock types from the older series, excluding those of spilites, amphibolites, and one intrusive porphyry, they seem to be reasonably representative of the bulk of the older (pre-Oligocene) rocks of the island. The sedimentary rocks, excluding the limestones, would probably show some increase in silica and alumina, but would otherwise be similar, as they are mainly directly derived from the volcanic rocks. These analyses have been plotted on a triangular diagram together with the Utuado analyses (Fig. 5). The two groups fall quite close together, the lavas being grouped slightly toward higher Fe, Mg, Ca, and lower SiO_2.

Table 2 shows for comparison the averages of the major constituents of the lavas and those of the Utuado rocks, and also of the latter excluding the gabbro, which is of minor importance areally. This comparison also demonstrates the increase in SiO_2 and decrease in Fe, Mg, and Ca, and it also shows a slight increase in Al_2O_3 and very slight increase in alkalies.

Since the relative quantities of the various rock types present cannot be estimated at this

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PLATE 2.—PHOTOMICROGRAPHS OF SELECTED ROCKS

FIGURES 1–3.—Series of three thin sections taken across contact between quartz porphyrite and quartz diorite at ADJ.NW47E40S. Note the increase in grain size of the groundmass. In Figures 2 and 3 the relic texture of the old groundmass can be seen in the cores of the large feldspar porphyroblasts. (×40)

FIGURE 4.—Quartz pyroxene hornfels, Locality JAY.SW7S46E. (×40)

FIGURE 5—Contact between hornfelsed xenolith and granodiorite. Locality JAY.NW7N35W. (×40)

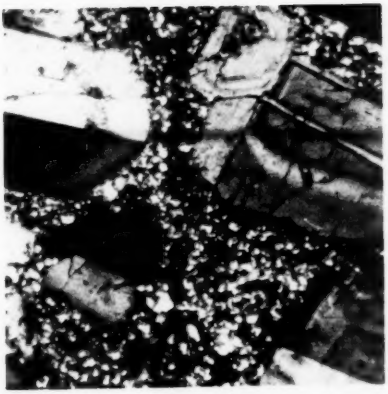

FIGURE 1

FIGURE 2

FIGURE 3

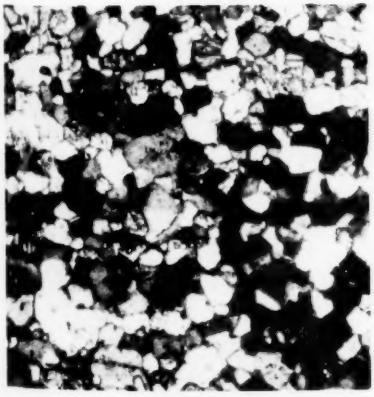

FIGURE 4

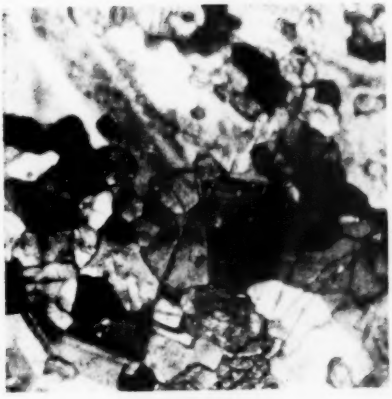

FIGURE 5

PHOTOMICROGRAPHS OF SELECTED ROCKS

FeO,
MgO

FIGU

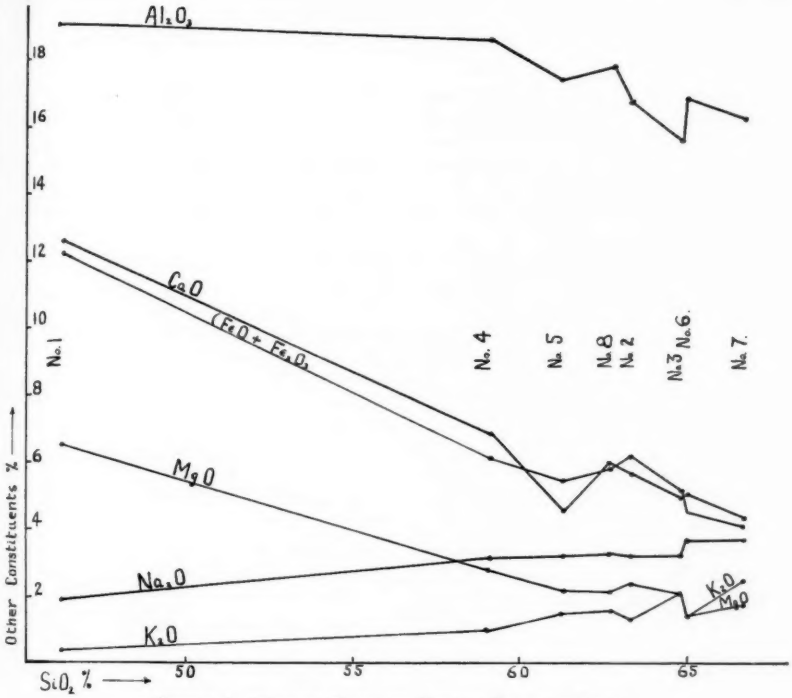

FIGURE 4.—HARKER DIAGRAM, UTUADO PLUTONIC ROCKS

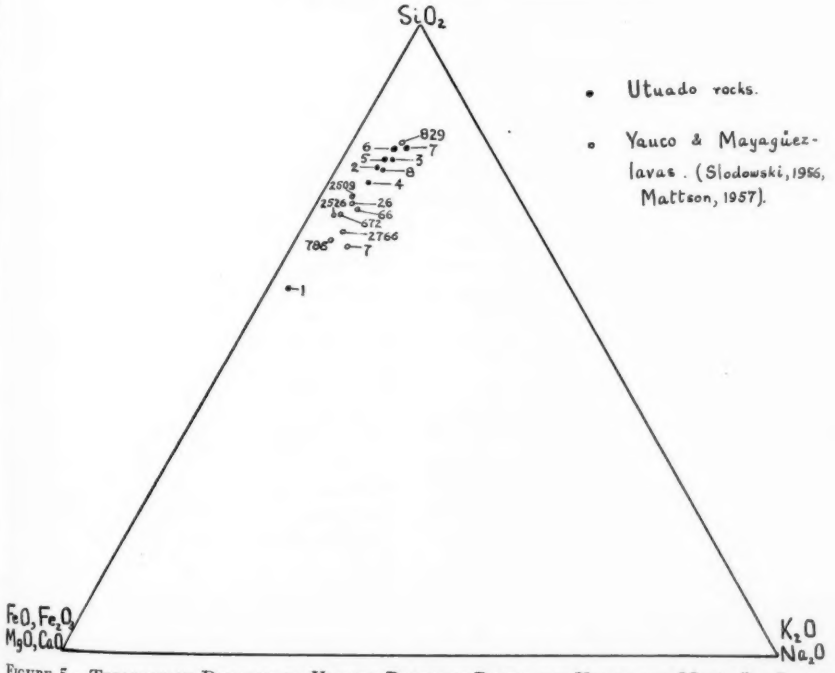

FIGURE 5.—TRIANGULAR DIAGRAM OF UTUADO PLUTONIC ROCKS AND YAUCO AND MAYAGÜEZ LAVAS FOR COMPARISON

stage of mapping, any far-reaching conclusion is precluded, but it can be said that the chemical evidence does not militate against the possibility of the diorites being formed by reconstitution of the pre-existing rocks.

One minor point should be mentioned. It will be noted that the analysis of the andesite from a relict pseudodike (No. 5) shows more than 5 per cent corundum in the norm. Holmes (1930, p. 436) has pointed out that this gives ground for suspecting a sedimentary origin for the rock. The quartz porphyrite (No. 6) also shows a high proportion of normative corundum—but here only 2.33. This may be due to weathering or alteration; nevertheless the two samples are fresh, and this excess of normative corundum may be significant.

All five lines of evidence as a whole present a fairly strong case for the contention that these dioritic rocks were formed by transformation of the preexisting rocks, which does not appear to have involved any major melting. The process seems to have been rather one of progressive recrystallization. In general it seems that the rocks went through a series of changes which can be represented by the various rock types present, *e.g.*, country rock-hornfels-quartz porphyrite-quartz diorite-granodiorite.

Table 2.—Averages of Major Constituents in Utuado Plutonic Rocks and of Yauco and Mayagüez Lavas

	SiO_2	Al_2O_3	FeO/Fe_2O_3	MgO	CaO	Na_2O	K_2O
Yauco & Mayagüez area lavas (9)*	56.19	15.11	7.56	5.15	7.42	2.95	1.14
Utuado plutonic rocks (8)	61.13	17.15	3.85	2.63	6.17	3.13	1.45
The same excluding gabbro (7)	63.26	16.89	2.66	2.08	5.27	3.31	1.61

* Slodowski (1956); Mattson (1957), Specimen Nos. P.R. 26, 672, 786, 829, 7, 66, 2766, 2509, 2526.

Age of the Pluton

The relative time of emplacement of the pluton can be estimated only within rather wide limits. It is certainly younger than the Upper Cretaceous (Maestrichtian) rocks it cuts, and it is almost certainly older than the middle Oligocene Lares limestone, which lies above it unconformably, completely unaltered.

It was previously thought (Meyerhoff, 1933) that all the rocks of the Older Series were Upper Cretaceous and that the major folding and plutonic intrusions took place in the very late Cretaceous or early Eocene. Now however, the picture is not so clear, since Kaye (1956), Slodowski (1956), and Mattson (1957) have demonstrated Paleocene rocks in the Older Series, and there is good reason to suspect the presence of Eocene rocks (Mattson 1957, p. 158). It seems that there have been not one, but at least three periods of deformation earlier than the pre-middle Oligocene unconformity.

Since the rocks in contact with the pluton contain poorly preserved fossils which indicate a Maestrichtian age, it can at least be said that the pluton is not older than Late Cretaceous. It could be Eocene. However, the relatively gently dipping (?) San Germán limestone outcrop south of the pluton suggests that the relatively greater deformation, including the anticline in which the pluton is emplaced, took place before deposition of the San Germán, *i.e.* in the late Maestrichtian. It does not, of course, necessarily follow that diorites were formed at the time of the greatest deformation. However, some fragmental rocks observed in reconnaissance outside the area of study, on Route 139 near Hacienda Carmelita, contain large pieces of quartz diorite. This suggests that the diorites had already been emplaced and exposed to erosion well before the end of the time of formation of the Older Series. At present then, the balance of evidence seems to favor a late Maestrichtian age for the pluton. Mitchell (1953) indicated a late Eocene age for these rocks but does not give the evidence upon which this statement is based. Kaye (1956, p. 120) has contended that the principal orogeny and plutonic intrusions are early Tertiary and that the importance of the late Cretaceous orogeny is discredited. This does not seem to be established. The author does not wish to minimize the orogenic activity in the early Tertiary, but he believes a considerable degree of tectonic deformation took place in the Late Cretaceous.

Whenever the diorites were emplaced,

whether approximately in the pre-San Germán interval or during Paleocene or early Eocene time, it seems hardly possible that they could have been formed at any great depth below the surface. Some vertical movements are necessarily implied by the unconformities, but in view of the general similarities of the rocks of the Cretaceous and of the Lower Tertiary—implying a general similarity of environment—it is inconceivable that these rocks could have been carried to depths in the crust notably greater than that implied by the thickness of the formations, that is not more than 2 or 3 miles at the utmost. This is a maximum estimate. According to Slodowski (1956, p. 88, 97) the San Germán formation attains a thickness of only 50 m in the Yauco area, and the Jicara formation above it only about 1000 m. Kaye (1956, p. 120) estimates more than 6000 feet of Lower Tertiary rocks in the Older Series around San Juan. This would imply a burial of little more than a mile.

Summary and Conclusions

This paper is a preliminary description of the Utuado plutonic complex in north-central Puerto Rico. These rocks which are of granodioritic and quartz dioritic composition (with minor gabbro) are emplaced in rocks of Maestrichtian age which mainly consist of volcanic rocks and volcanically derived sedimentary rocks. The structural, petrological, chemical, and field evidence point to the conclusion that these diorites were formed by reconstitution of the pre-existing rocks, thus supporting Ewing's (1953) hypothesis that these rocks formed by granitization of sediments in a trough. It should be pointed out, however, that the rocks from which they were derived bear little resemblance to those now found in the oceanic trenches.

The age of the pluton is not firmly established, but most evidence favors a Late Cretaceous age. It was probably formed close to the surface.

Considerably more detailed mapping, particularly in the areas surrounding the pluton, to connect up with areas where stratigraphical sequences have been established, is necessary before the age of the pluton and general geological history of the region can be firmly established.

Detailed study of the mineral changes involved in the formation of the plutonic rocks (particularly of the feldspars and hornblendes) is also desirable.

References Cited

Berkey, C. P., 1919, Geological introduction: N. Y. Acad. Sci., Sci. Survey Porto Rico and Virgin Islands, v. I, p. 11–29

Daly, R. A., 1933, Igneous rocks and the depths of the earth: New York, McGraw Hill Book Co. Inc., 598 p.

Ewing, M., 1953, Turbidity current transportation of sediments in the Atlantic (Abstract): Geol. Soc. America Bull., v. 64, p. 1419–1420

Goodspeed, G. E., 1955, Relict dikes and relict pseudo-dikes: Am. Jour. Sci., v. 253, p. 146–161

Holmes, A., 1930, Petrographic methods and calculations: London, T. Murphy & Co., 515 p.

Hubbard, B., 1923, The geology of the Lares district, Porto Rico: N. Y. Acad. Sci., Sci. Survey Porto Rico and Virgin Islands, v. II, pt. 1, p. 1–115

Kaye, C. A., 1956, Lower Tertiary of Puerto Rico: Am. Assoc. Petroleum Geologists Bull., v. 40, p. 108–121

—— 1957, Notes on the structural geology of Puerto Rico: Geol. Soc. America Bull., v. 68, p. 103–117

Mattson, P. H., 1957, Geology of the Mayagüez area, Puerto Rico: Princeton Univ. and the Commonwealth of Puerto Rico, Econ. Devel. Adm., 170 p.

McGuinness, C. L., 1948, Ground water resources of Puerto Rico: Puerto Rico Aqueduct and Sewer Service, 613 p.

Meyerhoff, H. A., 1927, Tertiary physiographic development of Puerto Rico and the Virgin Islands: Geol. Soc. America Bull., v. 38, p. 557–576

—— 1933, Geology of Puerto Rico: Univ. Puerto Rico Mon. ser. B., no. 1, 306 p.

Mitchell, R. C., 1953, New data regarding the dioritic rocks of the West Indies: Geol. en Mijn., new ser., Jaar. 15, p. 285–295

—— 1954, A survey of the geology of Puerto Rico: Univ. Puerto Rico, Agric. Expt. Sta. Tech. Paper 13, 167 p.

Roberts, R. C., 1942, Soil survey of Puerto Rico: U. S. Bur. Plant Industry, ser. 1936, no. 8, 503 p.

Rutten, L. M. R., 1939, The age of the quartz dioritic and granodioritic rocks of the West Indies: Geol. en Mijn., Jaar. 1, no. 5, p. 128–133

Slodowski, T. R., 1956, Geology of the Yauco area, Puerto Rico: Princeton Univ. and the Commonwealth of Puerto Rico, Econ. Dev. Adm., 130 p.

Turner, M. D., 1955, Geology of the San Sebastian area, Puerto Rico: 1st Carib. Geol. Conf., Antigua, B. W. I., 16 p. (mimeo.)

Weaver, J. D., 1957, Structure and Stratigraphy of the Copake quadrangle, New York: Geol Soc. America Bull., v. 68, p. 725–761

Zapp, A. D., Berquist, H. R., and Thomas, C. R., 1948, Tertiary geology of the coastal plains of Puerto Rico: U. S. Geol. Survey, Oil and Gas Inv., prelim. map 85

College of Agriculture and Mechanic Arts, University of Puerto Rico, Mayagüez, P. R.

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STRATIGRAPHY AND STRUCTURE OF THE NORTHERN SANGRE DE CRISTO MOUNTAINS, COLORADO

BY LINUS R. LITSEY

ABSTRACT

Precambrian metamorphic and igneous rocks, Paleozoic sedimentary rocks, and Tertiary igneous rocks are exposed in the northern Sangre de Cristo Mountains. The sedimentary rocks of Paleozoic age are most abundant and range from Ordovician through Permian. The pre-Pennsylvanian sedimentary rocks average about 1000 feet thick and are for the most part limestones and dolomites with lesser amounts of shale, sandstone, and quartzite. The Pennsylvanian and Permian rocks, more than 14,500 feet thick, are coarse clastic rocks deposited in the Colorado-New Mexico zeugogeosyncline. The Pennsylvanian rocks are poorly stratified and poorly sorted, drab sandstones and fine conglomerates; red sandstones and arkosic conglomerates predominate in the Permian rocks.

A tonalite stock of Tertiary age intrudes Pennsylvanian rocks in the southern part of the mapped area. Small dikes and sills, probably of Tertiary age, are present locally elsewhere in the area.

The northern Sangre de Cristo Range is a fault block uplifted along a concealed, high-angle fault that separates the range from the San Luis Valley to the west. The sedimentary rocks in the northern Sangre de Cristo Mountains strike generally north-northwest, nearly parallel to the trend of the range, and dip east 30° or more, forming the east flank of the Laramide Sawatch arch. Numerous high-angle strike faults are present. In general, faulting is more prominent than folding. All structures other than those of the Precambrian are probably Laramide or younger in age.

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Introduction

Location

The Sangre de Cristo Range, one of the most spectacular mountain ranges in Colorado, is geologically one of the least known ranges in the state. It extends from north-central New Mexico to south-central Colorado as a long, narrow, rugged range; in Colorado it is bound on the east by the Arkansas River Valley, Wet Mountain Valley, and Huerfano Park (Fig. 1), and on the west it is bounded by the broad, flat San Luis Valley. The northern termination of the range is at the Arkansas River near Salida, Colorado.

The fluvial topography is modified by alpine glacial features in the higher parts of the range. The crest of the range is sharply defined in most places, and peaks range from 12,500 feet to more than 14,000 feet in altitude. The east side of the range is drained by tributaries of the Arkansas River, but the water issuing from the west side sinks into coalescing alluvial fans along the east side of the San Luis Valley. Relief is about 5500 feet in the mapped area.

The area described in this report is in parts of Saguache, Custer, and Fremont counties, Colorado. The part west of the crest of the range is in Rio Grande National Forest; the part east of the crest is in San Isabel National Forest. From the east the area can be reached by gravel roads connecting with U. S. Highway 50, which follows the Arkansas River. Access to the western side is attained by gravel roads that lead east from U. S. Highway 285 in the San Luis Valley. The interior of the area is traversed only by trails.

Mapping of the southern two-thirds of the area described in this report was begun in 1951; about 6 months were spent in the field during the summers of 1951, 1952, and 1953. The remainder of the area was mapped by J. W. Rold and W. D. Siapno. Field mapping was done on U. S. Forest Service aerial photographs having a scale of approximately 1:20,000. Geologic data were transferred to an enlarged Forest Service drainage map.

Previous Work

The geology of the region was first examined by Endlich (1874; 1877), who made a cursory examination of the Sangre de Cristo Range, San Luis Valley, and adjacent regions for the Hayden Geographical and Geological Survey of the Territories. Bagg (1908) described copper deposits in the Pennsylvanian and Permian sandstones near Rito Alto Peak. Worcester (1919) investigated molybdenite claims near the same peak. Johnson (1929) discussed the stratigraphy of part of the range. Stone (1934) interpreted limonite deposits at the Orient mine as having originated from deep oxidation of iron carbonate minerals. Burbank and Goddard (1937) published a comprehensive report on the structural history of the Huerfano Park region, which is about 35 miles southeast of the area of this report. Many of their conclusions are applicable to the whole range.

More recently, J. R. Coash (1949, unpub. MS thesis, Univ. Colorado) described Pennsylvanian and Permian rocks in the Howard-Wellsville area; J. W. Rold (1950, unpub. MS thesis, Univ. Colorado) mapped and described the geology of the Wellsville area; and W. D. Siapno (1953, unpub. MS thesis, Univ. Colorado) mapped and described the geology of the Calcite area adjoining the Wellsville area to the south. Data from the papers by Rold and Siapno have been included in this report. P. Toulmin III (1953, unpub. MS thesis, Univ. Colorado) studied the petrography and petrology of the stock at Rito Alto Peak.

The only previous work that applies directly

to the area studied is a reconnaissance study by Gabelman (1952), which did not include detailed maps. He discussed the stratigraphy, major structural features, and Laramide and later structural history of a large part of the area in the Sangre de Cristo Range that is discussed in this report.

Acknowledgments

The writer wishes to express appreciation to Dr. John Chronic of the University of Colorado for criticism, encouragement, and direction of this work. The writer is indebted to William Siapno and John Rold for allowing their work in adjacent regions to be summarized in this paper. Ogden Tweto made many helpful suggestions on the organization and presentation of this material. Mark Melton and Allen Carey served as field assistants during the summers of 1951 and 1953, respectively.

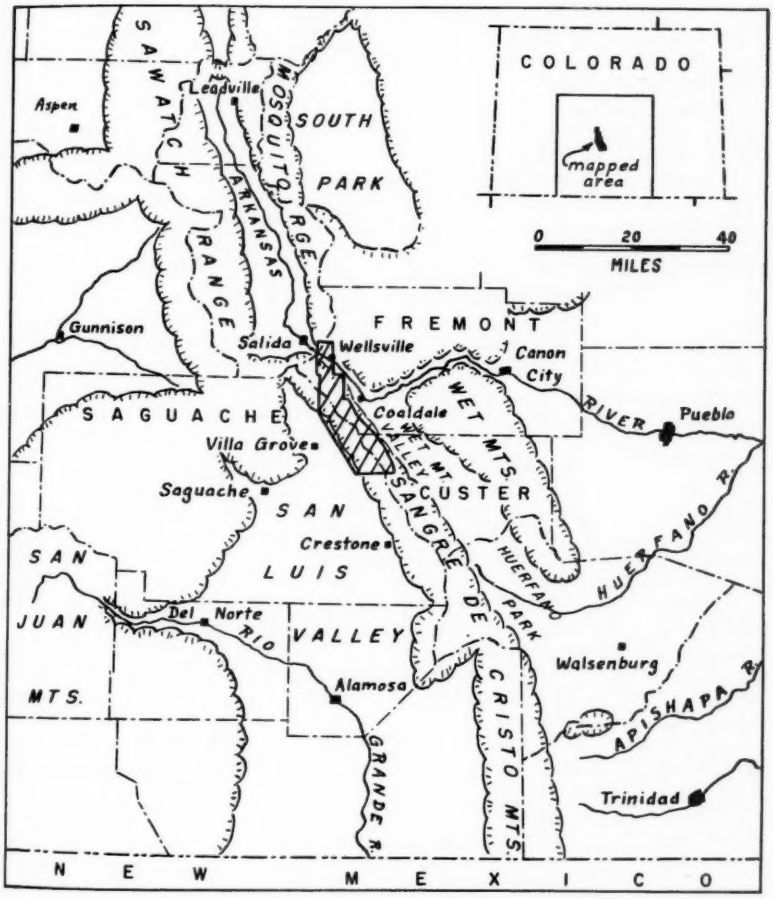

FIGURE 1.—INDEX MAP OF SOUTH-CENTRAL COLORADO

GEOLOGIC SETTING

Precambrian metamorphic and igneous rocks, Paleozoic sedimentary rocks, and Tertiary igneous rocks are exposed in the northern Sangre de Cristo Mountains. Sedimentary rocks of Mesozoic age were probably deposited in the area but they have since been removed by erosion. Surficial deposits, which include Quaternary gravels in the San Luis Valley, glacial and glaciofluvial deposits, and rock

streams in the mountains, were mapped in the southern two-thirds of the area but were not studied in detail. This report considers in detail only the sedimentary rocks of Paleozoic age and the structures that affect them.

North of Hayden Pass the crest of the Sangre de Cristo Range is carved in Precambrian rocks (Pl. 1). South of Hayden Pass the crest of the range is composed principally of Pennsylvanian and Permian sedimentary rocks that have been intruded locally by Tertiary igneous rocks. Precambrian rocks are exposed along the west side of the range as far south as the Orient mine and in the southwestern part of the mapped area south of Major Creek. The contact between the Paleozoic and Precambrian rocks served as a general western boundary for detailed mapping. Although Precambrian rocks are extensive in the northern end of the range, little time was devoted to their study.

The Precambrian rocks are composed of schist, fine-grained hornblende gneiss, quartz-biotite gneiss, and granite. In many places foliation in the metamorphic rocks is well developed and can be mistaken for bedding from a distance. The rocks are apparently folded into distorted isoclinal folds that trend generally north; the dips are steep or vertical (Gabelman, 1952, p. 1581). An erosional unconformity with less than 10 feet of local relief separates the Precambrian rocks from overlying sedimentary rocks. Weathering of the Precambrian rocks below this contact is shallow, not exceeding 5 feet in most places.

In the vicinity of Rito Alto Peak (T. 46 N., R. 11 E. in Pl. 1), the Pennsylvanian sedimentary rocks are intruded by the northernmost of three Tertiary stocks exposed along the crest of the range; *see* map of Colorado (U. S. Geological Survey, 1935). Endlich (1877) considered these stocks to be Precambrian granite and concluded that the range was anticlinal. Near Wellsville, the northern boundary for mapping was the contact of another Tertiary stock (J. W. Rold, 1950, unpub. MS thesis, Univ. Colorado). Numerous small dikes and sills, probably of Tertiary age, intrude rocks of all ages present.

The principal rock of the Rito Alto stock is a medium- to coarse-grained tonalite composed mainly of plagioclase, orthoclase, quartz, and biotite. The center of the tonalite body has a higher percentage of ferromagnesian minerals and a lower percentage of quartz than does the remainder. Minor rock types associated with the tonalite are porphyritic granite, quartz-rich granite, and an intrusive rhyolite.

A contact-metamorphic aureole is present around the stock, and the sediments have been recrystallized and silicified to varying degrees. On the east side of the stock where the effects of local metamorphism extend more than half a mile from the contact, epidote pervades the sandstone and is associated with small masses of garnet. Three linear zones of altered red rock, characterized by the presence of hematite, are present in and near the tonalite intrusive body. One zone, near Silver Lake, is entirely within the stock; the other two zones extend several hundred yards into the sedimentary rocks parallel to bedding. Within these altered zones the rocks are extensively argillized and sericitized, and the iron is oxidized to hematite (P. Toulmin III, 1953, unpub. MS thesis, Univ. Colorado).

Prospectors dug many prospect pits and small mines in the northern Sangre de Cristo Mountains, but their search for economic mineral deposits was for the most part without success. There are at least three different types of mineral deposits in the northern Sangre de Cristo Mountains, based on the metals contained. These are deposits that contain (1) iron; (2) lead, silver, copper, and gold; and (3) molybdenum. In general, the value of ore has been low, but small tonnages of rich ore have been produced sporadically. No mines were studied in detail, and no petrographic work was done on ore samples. Four districts discussed by Gabelman (1953) are within the area described in this report.

The largest mine is the Orient mine (Pl. 1) which supplied iron ore to the smelters of the Colorado Fuel and Iron Corporation in Pueblo, Colorado, intermittently from 1880 to 1933 (Gabelman, 1953, p. 196). The ore deposit is a replacement body in the lower part of the Leadville limestone. The primary ore, iron carbonate, was later deeply oxidized to limonite by meteoric waters (Stone, 1934).

Small veins (less than 1 foot thick) near Galena Peak contain argentiferous galena and chalcopyrite. These veins cut both Precambrian and lower Paleozoic rocks. Gold is present in some of these veins, and Butler (1949) reported tetrahedrite and sphalerite in veins near the Bushnell Lakes, 3 miles to the north. Similar veins are present in Steel Canyon in the lower Paleozoic sequence. Gangue minerals are chiefly quartz and calcite. The ore shoots are small and isolated.

In the Cloverdale mine east of Rito Alto Peak (Pl. 1) gold, copper, and silver were found in "gougy" material in minor faults. These faults were not detected at the surface. The host rock is sandstone of Permian (?) age.

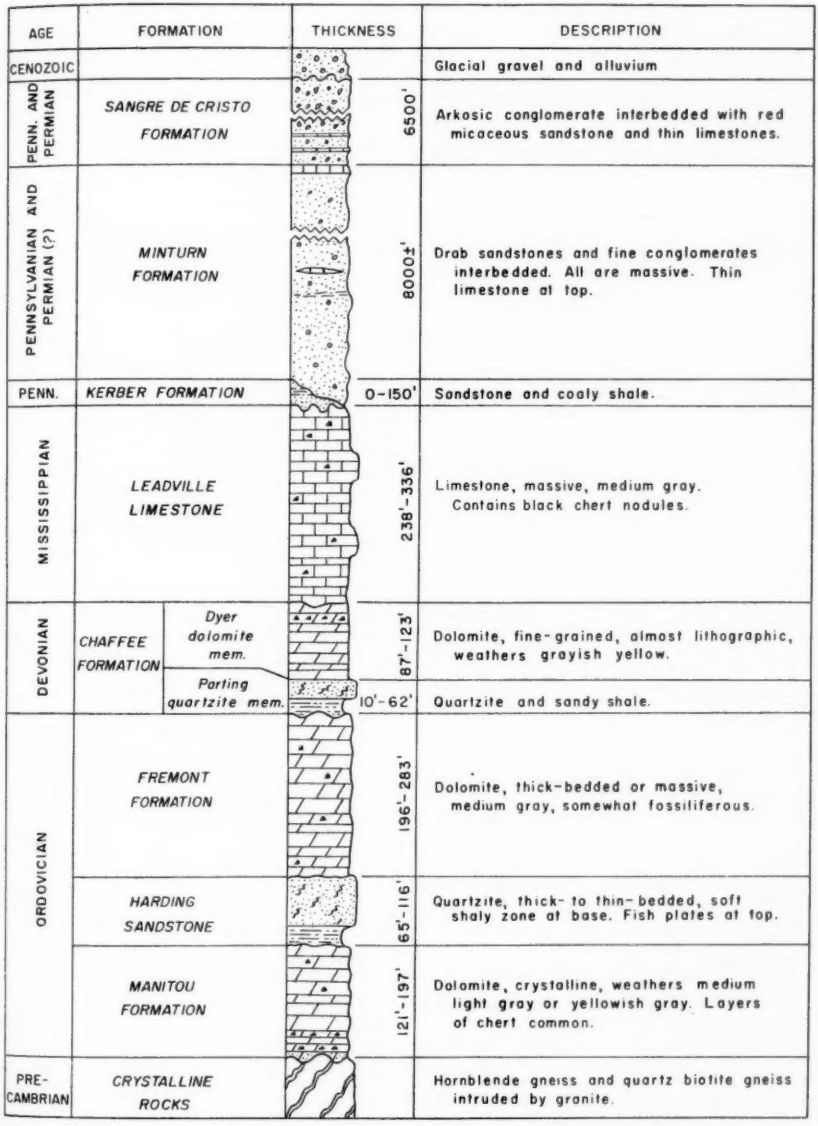

FIGURE 2.—COMPOSITE PALEOZOIC SECTION IN NORTHERN SANGRE DE CRISTO RANGE

Quartz veins in the Rito Alto stock near Silver Lake contain small crystals or fine-grained, structureless aggregates of molybdenite. The veins are about 1 foot wide and have a maximum length of a few hundred feet. Pyrite and chalcopyrite are associated with the molybdenite. Worcester (1919) examined claims on these veins and reported that selected samples assayed 3.47 per cent MoS_2, but that the whole vein probably would not average more than 1 per cent.

Probably all the mineralization in the area is hydrothermal and related to Tertiary igneous activity, although direct relationships cannot be proved in all places. Most of the ore, except

at the Orient mine, seems to occur as cavity fillings rather than as replacement bodies. Gabelman (1953, p. 190) offered several lines of evidence to suggest that most of the ores are low temperature epithermal deposits.

south of the trough in southern Colorado, New Mexico, and adjacent areas. The trough was intermittently inundated by seas during early and middle Paleozoic time; periodic minor uplifts resulted in disconformities between

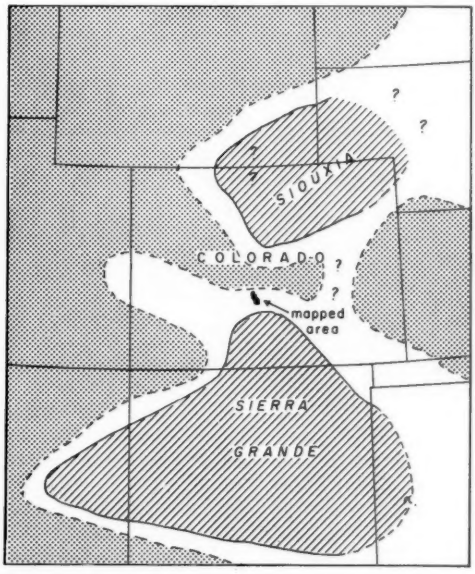

FIGURE 3.—LATE CAMBRIAN PALEOGEOGRAPHY

STRATIGRAPHY

Paleogeography

The sedimentary rocks in the northern Sangre de Cristo Range range from Ordovician through Permian (Fig. 2). The pre-Pennsylvanian units are thin and consist mainly of limestone and dolomite with lesser amounts of quartzite, sandstone, and shale. The lithology indicates stable shelf deposition. This area was near the southern edge of an east-west trough that extended across Colorado prior to the Pennsylvanian period. The paleogeographic map of the Late Cambrian (Fig. 3) illustrates the general position of the positive areas and areas of deposition throughout the pre-Pennsylvanian part of the Paleozoic. The positive area known as Siouxia was north of the trough, and the positive area of Sierra Grande was

most of the formations. The thicknesses of the pre-Pennsylvanian rocks measured in the northern Sangre de Cristo Mountains are shown in Table 1.

Pennsylvanian and Permian rocks were deposited as a thick series of clastic sediments in a zeugogeosyncline. The area of the northern Sangre de Cristo Mountains was a narrow part of this trough between two highlands of the Ancestral Rocky Mountains. These sedimentary rocks record the only major Paleozoic uplift in Colorado.

Figures 4–10 show the extent of Paleozoic rocks in central and southwestern Colorado. The thicknesses indicated on these maps were taken from the sections listed in Table 2. Only the formations that occur in the northern Sangre de Cristo Mountains are discussed in the text.

TABLE 1.—THICKNESSES OF PRE-PENNSYLVANIAN FORMATIONS IN THE NORTHERN SANGRE DE CRISTO MOUNTAINS

Area	Thickness of formations (in feet)*					
	Manitou formation	Harding sandstone	Fremont formation	Chaffee formation		Leadville limestone
				Parting quartzite member	Dyer dolomite member	
Wellsville†	197	65	229	54	92	238
Calcite**	.....	.....	196	62	87	266
Bushnell Lakes	179	84	.....	.....	.....	.....
Galena Peak	193	111	283	19	107	126††
South Piney Creek	.....	.....	.....	.....	.....	336
Steel Canyon	.....	.....	.....	15	105	.....
Black Canyon	121	116	249	10	123	.....

*Where thickness not shown, formation present but not measured.
†Measured by J. W. Rold (1950, unpub. MS thesis, Univ. Colorado).
**Measured by W. D. Siapno (1953, unpub. MS thesis, Univ. Colorado).
††Only lower part of Leadville measured.

TABLE 2.—MEASURED SECTIONS CITED ON DISTRIBUTION MAPS

Map Reference Number	Locality	Author	Map Reference Number	Locality	Author
1	McCoy	Donner (1949)	28	Trout Creek	Johnson (1934)
2	Dotsero	Bassett (1939)	29	Weston Pass	Johnson (1934)
3	Glenwood Springs	Vanderwilt and Fuller (1935)	30	Leadville	Behre (1939)
4	East Lake Creek	Johnson (1944)	31	Pando	Tweto (1949)
5	Brush Creek	Johnson (1944)	32	Gilman	Johnson (1944)
6	Wood's Lake	Johnson (1944)	33	Kokomo	Koschman & Wells (1946)
7	Frying Pan River	Johnson (1944)			
8	Aspen Mountain	Johnson (1544)	34	Alma	Singewald & Butler (1932)
9	Tourtelette Park	Johnson (1944)			
10	Ashcroft	Johnson (1944)	35	Missouri Gulch	Maher (1950)
11	Taylor Pass	Johnson (1944)	36	Gove Canyon	Maher (1950)
12	Mt. Milton	Johnson (1944)	37	Williams Canyon	Maher (1950)
13	North Italian Mt.	Johnson (1944)	38	Deadman's Canyon	Maher (1950)
14	Yule Creek	Vanderwilt and Fuller (1935)	39	Phantom Canyon	Maher (1950)
15	Cement Creek No. 2	Johnson (1944)	40	Oil Creek	Maher (1950)
16	Cement Creek No. 1	Johnson (1944)	41	Canon City	Maher (1950)
17	Deadman's Creek	Johnson (1944)	42	Specimen Hill	Maher (1950)
18	Cross Mountain	Johnson (1944)	43	South Hardscrabble Creek	Maher (1950)
19	Fossil Ridge	Johnson (1944)			
20	Tincup	Johnson (1944)	44	Beulah	Maher (1950)
21	Round Mountain	Johnson (1944)	45	Ouray	Burbank (1930)
22	Monarch-Tomichi	Crawford (1913)	46	Rico	Cross and Ransome (1905)
23	Kerber Creek	Burbank (1932)			
24	Black Canyon	Litsey (Table 1)	47	Engineer Mountain	Cross (1910)
25	Galena Peak	Litsey (Table 1)	48	Needle Mountain	Cross *et al.* (1905)
26	Howard	J. R. Coash (1949*)	49	Animas River	Bass (1944)
27	Wellsville	J. W. Rold (1950*)			

*Masters thesis, Univ. Colorado (unpublished).

Cambrian System

No sedimentary rocks known definitely to be Cambrian in age are present in the northern Sangre de Cristo Range. The Sawatch quartzite of Late Cambrian age has been reported in the Wellsville area along the Arkansas River just east of the Chaffee-Fremont county line by Brainerd, Baldwin, and Keyte (1930) and Johnson (1945, p. 8), but J. W. Rold (1950, unpub. MS thesis, Univ. Colorado) reported this same unit as Precambrian because: (1)

the clear-cut contact that in most places is between the Sawatch quartzite and the Precambrian rocks is not present; (2) rocks almost identical with the "Cambrian rocks" crop out several hundred feet below Precambrian rocks

Manitou formation, the Harding sandstone, and the Fremont formation. Unconformities separate these three formations from each other, from younger rocks above, and, in places, from Cambrian rocks below. In some

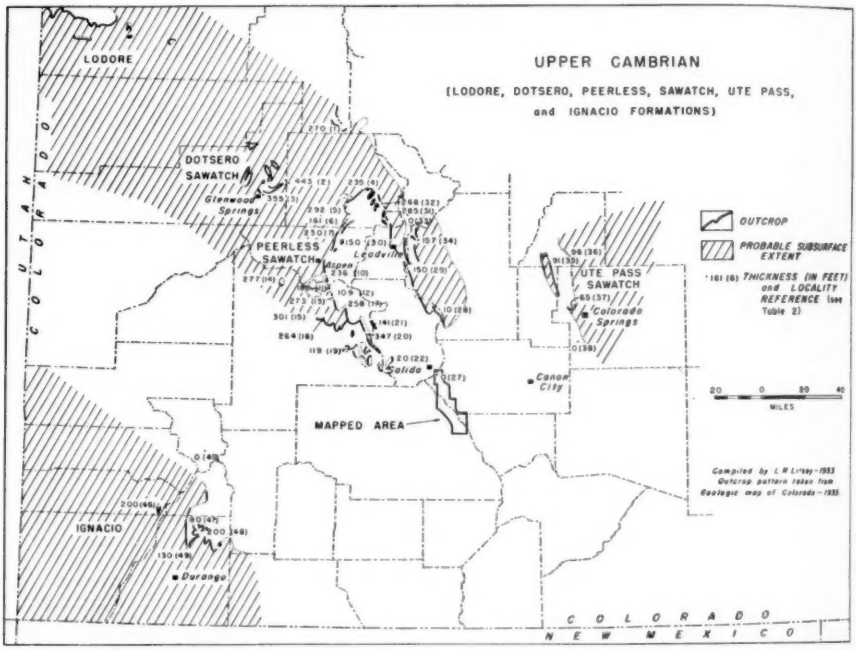

FIGURE 4.—THICKNESS AND EXTENT OF UPPER CAMBRIAN SEDIMENTARY ROCKS

in Spring Gulch; (3) north of the Arkansas River several quartz veins transect the "Sawatch" but are truncated by the erosion surface at the base of the Manitou formation. No evidence was found to suggest an age other than Precambrian for these veins; (4) the "Cambrian rocks" are as intensely metamorphosed as the underlying Precambrian.

The nearest rocks of unquestioned Cambrian age are at Trout Creek (Johnson, 1934) about 20 miles to the north, and in the Monarch-Tomichi district (Crawford, 1913), about 20 miles to the northwest. Cambrian rocks have not been identified in the Wet Mountains. Figure 4 shows the extent and thickness of the Upper Cambrian sequence.

Ordovician System

General statement.—Ordovician formations in Colorado are, from oldest to youngest: the

places in Colorado, Ordovician rocks rest on Precambrian rocks. Although the Harding sandstone and Fremont formation do not extend as far north as the Manitou formation, they probably once covered a much larger area than their present distribution indicates.

Manitou formation.—The Manitou formation was originally named the Manitou limestone by Cross (1894) for excellent exposures in Manitou Park and Manitou Springs, Colorado. As it was in doubt which of these was the type locality, Brainerd *et al.* (1933, p. 380) redefined the formation from exposures in Williams Canyon near Manitou Springs. Maher (1951) assigned the lowest few feet of the Manitou in the Manitou Springs area to the Late Cambrian (?) and named it the Ute Pass dolomite, restricting the Manitou to the overlying beds. Bass and Northrop (1953, p. 905) used the name Manitou formation in the White River Plateau area rather than Manitou limestone,

because the unit contained several rock types. Most of this unit is dolomite in the Sangre de Cristo Mountains, but the name Manitou formation is used, because several rock types are present in other parts of Colorado. Figure 5 shows the distribution of the Manitou formation as it is now defined.

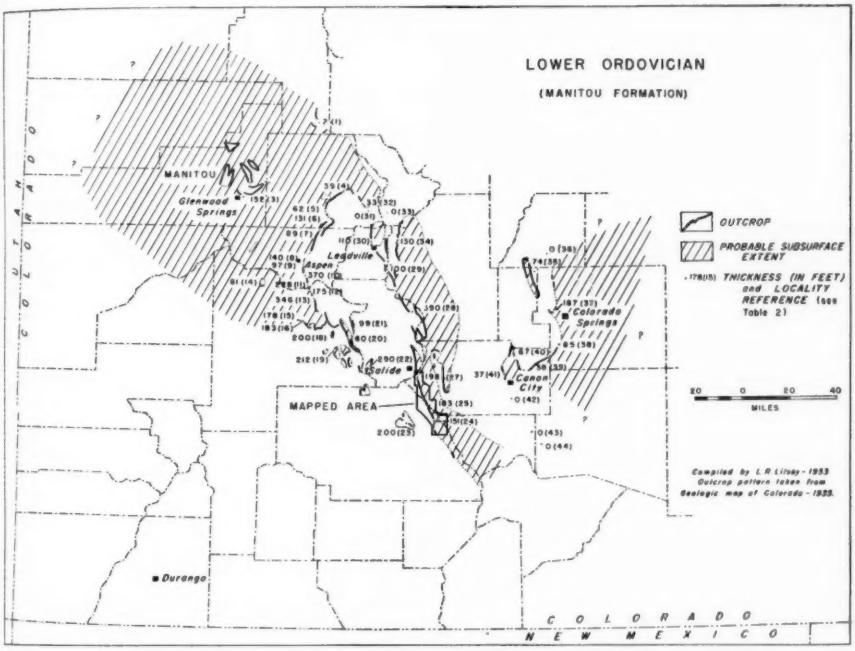

FIGURE 5.—THICKNESS AND EXTENT OF LOWER ORDOVICIAN SEDIMENTARY ROCKS

The Manitou formation is the most widespread Ordovician unit in Colorado. In most areas it overlies Cambrian rocks, in places with gradational contact, but along the Front Range from Colorado Springs to Canon City, and in the northern end of the Sangre de Cristo Range, the Manitou extends beyond the limits of the Cambrian rocks and rests on the Precambrian.

The Manitou ranges from a few feet to approximately 400 feet thick along Trout Creek and in the Ashcroft area near Aspen (Fig. 5). In the northern Sangre de Cristo Mountains thicknesses range from 121 to 197 feet, which may reflect differences in amount of original deposition. Radical variations in thickness, such as those northwest of Salida in the Sawatch Range (Fig. 5), probably indicate erosion prior to deposition of younger beds.

Typically the Manitou consists of thin-bedded, somewhat siliceous dolomite. In some areas the formation is predominantly limestone or dolomitic limestone; locally, there are sands or shales at the base. In the northern Sangre de Cristo Mountains the Manitou is almost entirely fine- to medium-grained dolomite. Characteristically it is medium-bedded. Beds range from a few inches to 2 feet thick. For the most part fresh surfaces are medium gray to medium light gray and weathered surfaces are light gray to yellowish gray. Rhombs of dolomite weather out to form sandpaperlike surfaces; at some horizons the rough surface is due to silt or fine sand. Layers of light-gray to pink chert, generally 1–3 inches thick, are common in the lower 20 feet of the formation and are present in lesser amounts throughout the entire thickness.

The Manitou lies unconformably on a surface of low relief cut on Precambrian rocks. At no place in the area has shale, sandstone, or conglomerate been observed at the base of the Manitou, but the basal dolomite (1–2 feet thick) commonly contains a small amount of clastic material with sparse grains as much as a quarter of an inch in diameter.

Fossils are rare and poorly preserved in the Manitou formation, but some have been found in the type section. No fossils were found in the area of this report. Johnson (1945, p. 18) has compiled a composite fauna from the Manitou that indicates it is Early Ordovician.

Harding sandstone.—The Harding sandstone was originally described and named by Walcott (1892, p. 155) for exposures in and near the Harding sandstone quarry, about 1 mile northwest of the state penitentiary at Canon City, Colorado. The Harding sandstone is 86 feet thick in the type section; it has a maximum thickness of 151 feet in Priest Canyon, about 5 miles north of Canon City (Brainerd *et al.*, 1933, p. 384). Figure 6 shows the present distribution of the Harding sandstone.

Throughout the northern part of the Mosquito and Sawatch ranges the Harding is missing, except in the Pando area, where Tweto (1949, p. 166) reported a maximum thickness of 50 feet. In the Pando area the sandstone overlies the Peerless formation and underlies the Parting quartzite member of the Chaffee formation; it is separated from formations above and below by unconformities. The thickness of this unit increases to the north and west. If Tweto's correlation is correct, the Harding sandstone of the Pando area is probably an outlying remnant that survived Silurian and Early Devonian erosion. The Harding thickens southward from a knife edge a few miles north of Salida to 116 feet thick in the northern Sangre de Cristo Range.

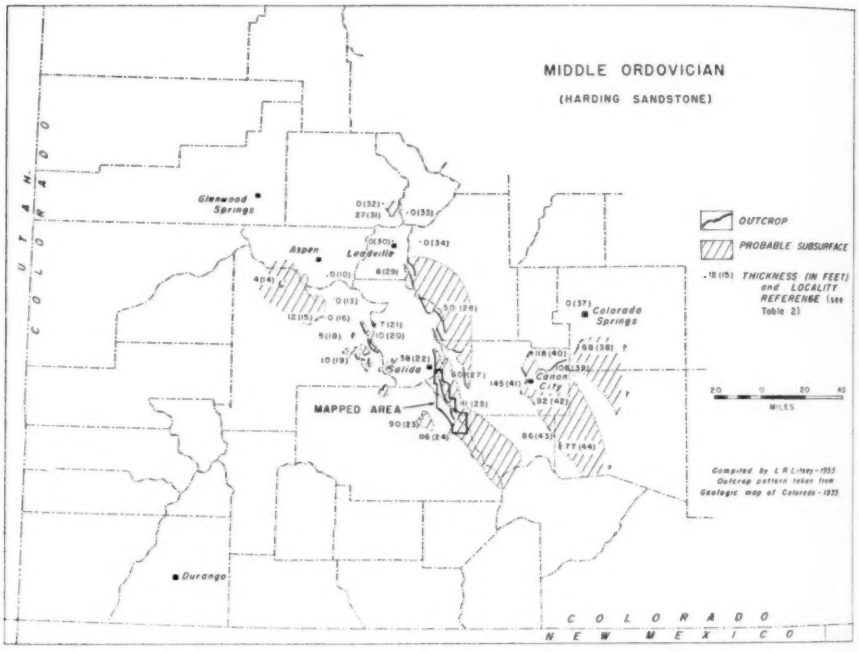

FIGURE 6.—THICKNESS AND EXTENT OF MIDDLE ORDOVICIAN SEDIMENTARY ROCKS

In the type locality the Harding sandstone consists of fine-grained sandstones with some varicolored shales and dolomitic limestones; the formation becomes more quartzitic to the west. In the northern Sangre de Cristo Mountains the Harding is characterized by a soft basal zone composed of grayish-red or green shale, sandy shale, and thin-bedded sandstones. This zone ranges from about 15 to 30 feet thick and is overlain by about 90 feet of resistant beds with individual beds 1.5–6 feet thick of light-colored, fine-grained quartzite. Thin-bedded quartzite and thin, shaly partings are present in places in the resistant zone. The quartzite is composed of well-rounded, well-sorted, pure, quartz sand. The fresh surface is light gray, but the weathered surface is commonly dark brown or reddish brown. The

resistant quartzite forms cliffs and prominent ridges. Small rounded ridges, as much as 3 inches long, on many bedding surfaces may be coprolites or molds of coprolites.

agree that it is probably Middle Ordovician (Walcott, 1892, p. 159; Kirk, 1930, p. 463; Flower, 1952, p. 505).

Fremont formation.—The Fremont formation

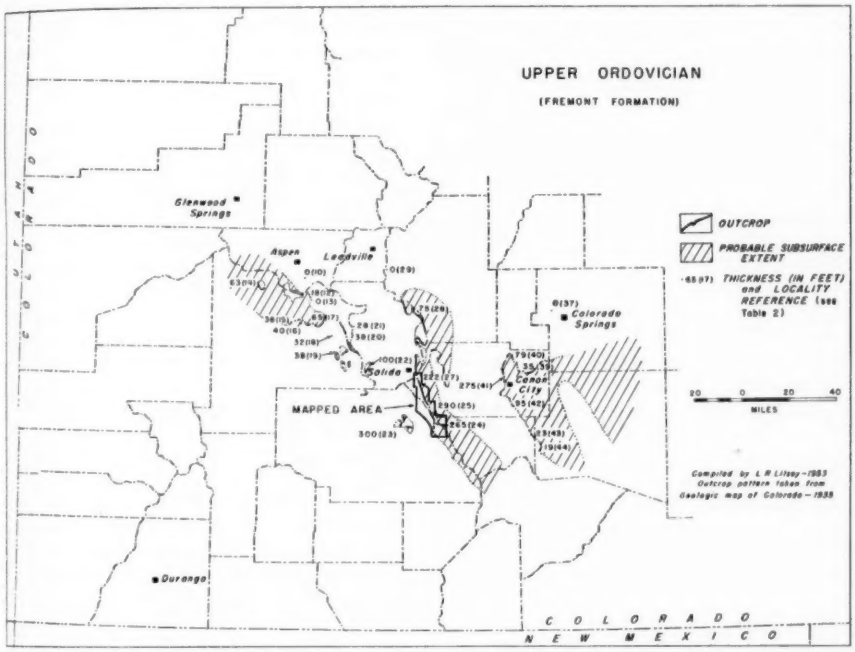

FIGURE 7.—THICKNESS AND EXTENT OF UPPER ORDOVICIAN SEDIMENTARY ROCKS

There is no evidence of an angular unconformity between the Manitou formation and the Harding sandstone, but a disconformity is indicated by the varying thickness of the Manitou and the sharp change in lithology at the contact. The contact is clearly exposed in only a few places because the soft basal beds of the Harding are easily weathered.

The Harding sandstone is only sparingly fossiliferous, and the few fossils present are generally poorly preserved. It has a limited, dominantly molluscan fauna; but, according to Sweet (1954, p. 291), there is a varied, well-preserved conodont fauna that seems to be useful in correlation. The only fossils seen in the area were fish plates, which were found in a zone only a few inches thick near the top of the formation. Fragmentary fish remains in the Harding are among the oldest known indications of vertebrate life.

The exact age of the Harding has not been determined with certainty, but several authors

was first called the Fremont limestone by Walcott (1892, p. 157) from exposures near Canon City, Fremont County, Colorado. As this unit is principally dolomite in the Sangre de Cristo Mountains and is a limestone in some areas, the name Fremont formation is used in this report. The distribution of the Fremont both in outcrop and subsurface extent (Fig. 7) is similar to that of the Harding sandstone. The formation is absent in the northern ends of the Mosquito and Sawatch ranges, but it appears between Weston Pass and Trout Creek in the Mosquito Range and thickens southward to a maximum of 300 feet at Kerber Creek.

The Fremont at the type locality is a massive limestone with arenaceous and argillaceous beds, but in other areas it is dolomite or dolomitic limestone. In the northern Sangre de Cristo Mountains the Fremont is almost entirely fine-grained, crystalline dolomite that is light gray and massively bedded. It weathers to a rough, jagged surface with numerous

spines and ridges that are separated by cup-shaped depressions 1.5–4 inches in diameter. Some chert is present in nodules and layers. According to Sweet (1954, p. 294), two lithologically and faunally distinct members of the Fremont formation can be recognized in Colorado, a lower massive dolomite member and his Priest Canyon member above. No lithological subdivision was possible in the northern Sangre de Cristo Mountains.

In most places the contact between the Harding sandstone and Fremont formation is covered, but evidence indicates that it is a disconformity. Near Galena Peak the contact is slightly undulatory. The basal bed of the overlying Fremont formation is a dolomitic sandstone bed 4 feet thick and may have resulted from reworking of the Harding.

The Fremont formation is fossiliferous, but fossil preservation is poor. In the area of investigation numerous horn corals were found, and the chain coral *Halysites* is fairly common. In addition, there are a few poorly preserved brachiopods, and near Middle Bushnell Lake several fragments of a straight cephalopod with a diameter of 4 to 5 inches were found. Walcott (1892, p. 163) concluded that the Fremont fauna indicated a late Trenton age. Johnson (1945, p. 26–29) lists the fauna and states that it indicates a Late Ordovician (Richmond) age.

Silurian System

No rocks of Silurian age have been found in Colorado. It seems probable that this area was low but positive and was undergoing weathering or mild erosion throughout the period. If any Silurian deposits were laid down, they have since been removed by erosion or have not been exposed.

Devonian System

General statement.—Sedimentary rocks of Late Devonian age are exposed along the flanks of mountain ranges from Glenwood Springs in central Colorado to the northern Sangre de Cristo Mountains. No older Devonian rocks have been found in Colorado. Devonian formation names in common use in Colorado are the Chaffee and Elbert formations and the Ouray limestone. The Williams Canyon limestone, which was previously considered to be Devonian, has recently been correlated with Mississippian strata in eastern Colorado (Maher, 1950). Figure 8 shows the present extent of Upper Devonian strata in Colorado

The oldest Devonian sedimentary rocks are clastic rocks which lie disconformably on older beds. In southwestern Colorado the clastic rocks are shale with minor sandstone and quartzite; in central Colorado sandstone and quartzite make up a larger proportion. In southwestern and central Colorado these clastic rocks are overlain by limestone or dolomite, and in some places the contact is gradational.

Chaffee formation.—The name Chaffee formation, from Chaffee County, Colorado, was first applied to all Devonian rocks of central Colorado by Kirk (1931, p. 230). In most of its outcrops the Chaffee can be divided into a lower sandstone or quartzite called the Parting quartzite member and an upper limestone or dolomite called the Dyer dolomite member.

PARTING QUARTZITE MEMBER: Emmons (1882) first applied the term Parting to the white quartzite, 10–70 feet thick, which separates the Blue limestone (Leadville and Dyer) from the underlying White limestone (Manitou) in the Leadville district. The type section is at the Parting spur of Dyer Mountain east of Leadville.

The Parting is primarily a quartzite near Leadville and farther north, but to the south and southwest the member becomes more shaly and calcareous and is lithologically similar to the Elbert formation of southwestern Colorado. In the northern Sangre de Cristo Range the Parting quartzite member commonly consists of red or olive-gray shale or dolomitic shale at the base overlain by conglomeratic quartzite. J. W. Rold (1950, unpub. MS thesis, Univ. Colorado) distinguished an Elbert facies of argillaceous dolomitic limestone at the base of the Parting near Wellsville. The change to the southwest toward lithology similar to the Elbert suggests that these two units were once a single stratigraphic unit that was continuous across the Uncompahgre highland prior to Pennsylvanian erosion (see Fig. 11).

The Parting varies in thickness; a maximum of about 75 feet is attained along Brush Creek in the Sawatch Range. The Parting is 10–62 feet thick in most places in the northern Sangre de Cristo Mountains, but in a small area near the Bushnell Lakes it is absent. It is an excellent stratigraphic marker.

The disconformity at the base of the Parting quartzite member is the most distinct break in deposition in the pre-Pennsylvanian strata of the northern Sangre de Cristo Range. At Galena Peak the basal bed is a conglomerate containing cobbles of the underlying Fremont dolomite as

much as 6 inches in diameter in a matrix of coarse quartz sand. In Steel Canyon the upper 14 feet of the Fremont is weathered. Here the Parting consists of 5 feet of grayish-red shale at the base, overlain in turn by 1½ feet of fine-grained sandstone and 8 feet of coarse-grained, massive quartzite.

According to Johnson (1945, p. 32), no fossils except fragmentary fish remains have been found in the Parting quartzite member. Because it is apparently correlative with the Elbert formation, it has been considered to be Late Devonian.

DYER DOLOMITE MEMBER: The Dyer dolomite member of the Chaffee formation was named by Behre (1932) from its typical exposure on West Dyer and Dyer mountains, 5 miles east of Leadville. It was formerly included with the Leadville limestone in the Blue limestone, but fossils collected by Kirk (1931, p. 227) established its age as Devonian.

The Dyer dolomite member varies considerably in thickness, and generally is thickest in the Sawatch Range; the maximum thickness known is about 200 feet near Aspen, Colorado (Fig. 1). In the northern Sangre de Cristo Mountains the Dyer ranges from 87 to 123 feet thick. According to Johnson (1945, p. 38):

"Field evidence clearly indicates that the differences in thickness of the Dyer are largely, or entirely, the result of erosion during the time interval preceding the deposition of Mississippian sediments."

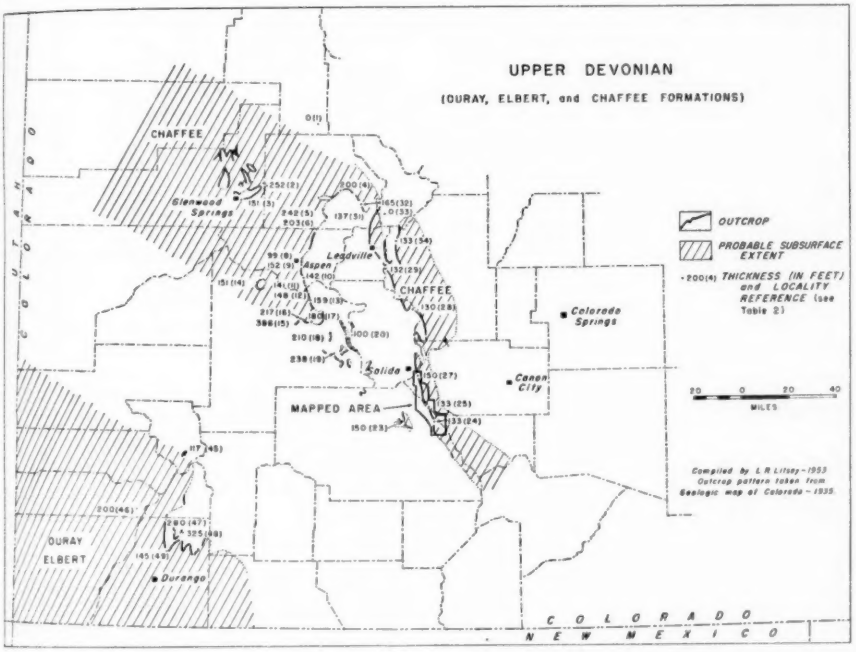

FIGURE 8.—THICKNESS AND EXTENT OF UPPER DEVONIAN SEDIMENTARY ROCKS

The Dyer is typically a dense, fine-grained, dolomitic limestone or dolomite; it is gray or bluish gray on a fresh surface but brown or yellow on a weathered surface. Thin shale partings are abundant near the base in some localities.

In the northern Sangre de Cristo Mountains, the Dyer is mostly dolomite with some intercalated beds of limestone. In general it has a lithographic texture and breaks with a subconchoidal fracture. J. W. Rold (1950, unpub. MS thesis, Univ. Colorado) and W. D. Siapno (1953, unpub. MS thesis, Univ. Colorado) have noted a zone containing "wire-like" chert near the top of the section. This chert occurs without regard to bedding, making the rock very tough and giving it a rough weathered

surface. This cherty zone was observed at Galena Peak, but it could not be distinguished in sections farther south.

In general, the contact between the Dyer and

for Mississippian strata in Colorado, except for the southeastern part of the state, are the Leadville, Madison, Williams Canyon, Hardscrabble, and Beulah limestones (Fig. 9). The

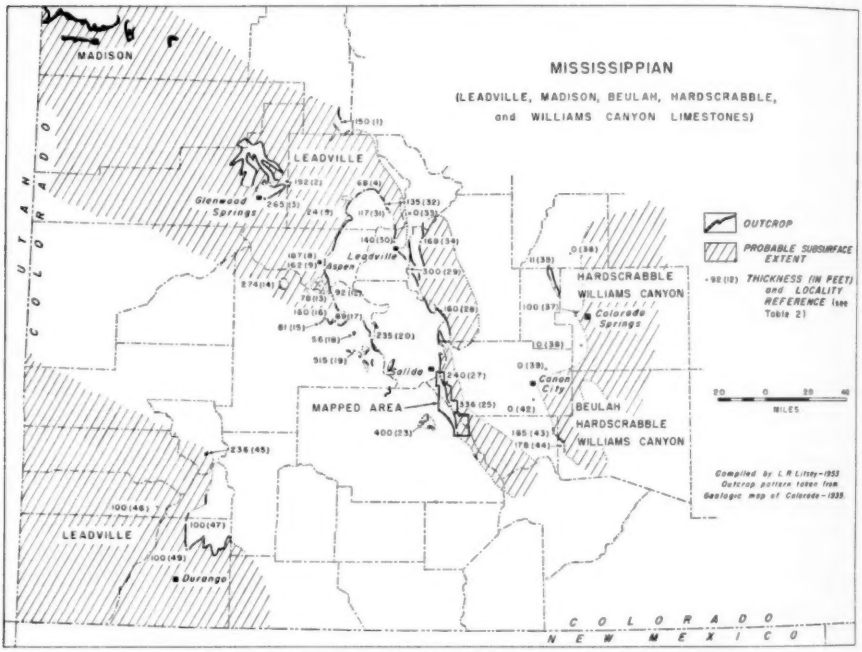

FIGURE 9.—THICKNESS AND EXTENT OF MISSISSIPPIAN SEDIMENTARY ROCKS

Parting members of the Chaffee formation is gradational, but, locally, a slight disconformity may exist. At Galena Peak the contact appears gradational, but in Steel Canyon it is slightly undulatory and lacks any gradational zone.

The Dyer is essentially nonfossiliferous in the Sangre de Cristo Range; however, it is fossiliferous in some parts of central Colorado. Johnson (1945, p. 36) lists the Dyer fauna and states that it correlates with that of the Ouray limestone in southwestern Colorado. The Dyer is Late Devonian.

Mississippian System

General statement.—Sedimentary rocks of Mississippian age are found throughout central and southwestern Colorado and extend eastward from Canon City under the great plains. These rocks probably once covered a much larger area than their present distribution indicates. Formation names in current usage name Millsap formation was used formerly for the Mississippian rocks east of the Front Range and Wet Mountains.

Mississippian sedimentary rocks are dominantly limestone, with some shale, thin sandstones, and limestone conglomerates at the base. The limestones are mostly medium dark gray and massively bedded. Thicknesses range from a few feet to more than 500 feet.

Leadville limestone.—The name Leadville limestone was originally applied to beds between the Manitou formation of Ordovician age and Pennsylvanian rocks in the Anthracite-Crested Butte region by Eldridge (1894). The name was derived from the Leadville mining district, where this sequence is the principal ore bearer. The upper part of the Leadville was assigned to the Mississippian system on fossil determinations by G. H. Girty (in Emmons *et al.*, 1927, p. 27). When the lower part was found to be Devonian, Kirk (1931, p. 227) restricted the name Leadville limestone to

those beds of Mississippian age. Kirk also suggested that the Mississippian part of the Ouray limestone of southwestern Colorado be called Leadville and that the name Ouray should be restricted to the Devonian part. This suggestion has been followed in the literature.

The great range in thickness of the Leadville limetone in central Colorado (Fig. 9) is primarily the result of erosion prior to deposition of the Pennsylvanian sediments. In the Sawatch Range the thickness ranges from 24 feet to a maximum of about 515 feet at Fossil Ridge. Measured thicknesses in the northern Sangre de Cristo Mountains range from 238 to 336 feet, but in many places measurements could not be made because the contact with Pennsylvanian beds was not exposed or because the section was faulted. Mississippian rocks are absent in the Paleozoic outlier about 10 miles east of Calcite on the east side of the Arkansas River (J. R. Butler, 1954, unpub. MS thesis, Univ. Colorado).

At the northern ends of the Sawatch and Mosquito ranges the Leadville is dolomite or dolomitic limestone, but it becomes less dolomitic to the south and southwest. Black chert nodules are common. Near Glenwood Springs, Colorado, a thick, massive, fine-textured, oölitic member occurs at the top of the formation.

The Leadville in the Sangre de Cristo Mountains is almost entirely a fine-grained, dark to medium-gray, massive, crystalline limestone. At some places it contains a few dolomite beds and, at several horizons, it contains limestone breccias of sedimentary origin. Near the Bushnell Lakes thin sandstone beds are present in the upper part of the Leadville; their lateral extent is apparently limited. Commonly two zones which appear to be more resistant than the rest form ridges near the base and top of the Leadville outcrop.

A thin, persistent sandstone and limestone conglomerate, which in most places does not exceed 20 feet in thickness, is present at the base of the Mississippian section in the Leadville district and adjacent areas. This clastic basal unit has been named the Gilman sandstone member of the Leadville limestone by Lovering and Tweto (1944) for exposures in the Minturn quadrangle. J. W. Rold (1950, unpub. MS thesis, Univ. Colorado) and W. D. Siapno (1953, unpub. MS thesis, Univ. Colorado) noted this member near Wellsville and Calcite, respectively. At Galena Peak the basal contact of the Leadville is exposed, but no sandstone or limestone conglomerate is present. Gabelman (1952, p. 1585) reported 5 feet of sandstone at the Orient mine.

One of the most interesting unconformities in Colorado is at the top of the Leadville limestone. Sinkholes and caverns filled with Pennsylvanian detritus indicate a pre-Pennsylvanian karst topography in many areas (Donner, 1949, and others). A red weathered zone containing Mississippian chert pebbles occurs at the base of the Pennsylvanian sequence in many places. In Lime Canyon a channel in the top of the Leadville about 10 feet deep and 30 feet wide is filled with bedded Pennsylvanian sandstone. In most places in the northern Sangre de Cristo Mountains the top of the Leadville is poorly exposed, and no other channels were observed.

Fossils are rare in the Leadville in the northern Sangre de Cristo Mountains, although a few poorly preserved brachiopods and horn corals were found. Johnson (1945, p. 52) and some earlier workers concluded that the Leadville limestone is Early Mississippian (Kinderhook and early Osage) in age. K. E. Scott (1954, unpub. MS thesis, Univ. Colorado) discovered a zone of brachiopods similar to those in the Madison limestone above the middle of the Leadville limestone in the northwestern part of the Sawatch Range. He concluded that the lower part of the Leadville of that area is Kinderhook and Osage in age and is equivalent to the Madison limestone of the type section. He also concluded that outliers of the Meramec series are locally present at the top of the Leadville because of the presence of the horn corals *Caninia coloradoensis* Scott ms. and *Triplophyllites radiatus* Scott ms., and a foraminifer, *Plectogyra* sp. *Caninia coloradoensis* is a newly found species that is intermediate in evolutionary development between corals that are dated as Osage and those dated as Chester. Scott considered this upper part of the Leadville equivalent to the lower part of the Brazer limestone. Maher (1950) correlated the Mississippian beds that crop out along the east side of the Front Range and Wet Mountains with an Upper Mississippian formation of the Meramec series of eastern Colorado and western Kansas.

Pennsylvanian and Permian Systems

General statement.—Pennsylvanian and Permian sedimentary rocks are exposed over a greater area in Colorado than the earlier Paleozoic rocks combined. Broad outcrops occur in a belt extending from the eastern end of the

Uinta Mountains, Utah, through the Sawatch, Mosquito, and Sangre de Cristo ranges in Colorado into New Mexico (Fig. 10). Numerous outcrops are present in southwestern Colorado, Sierra Grande uplift as a Pennsylvanian positive area which extended southeastward from the southern end of the Front Range highland.

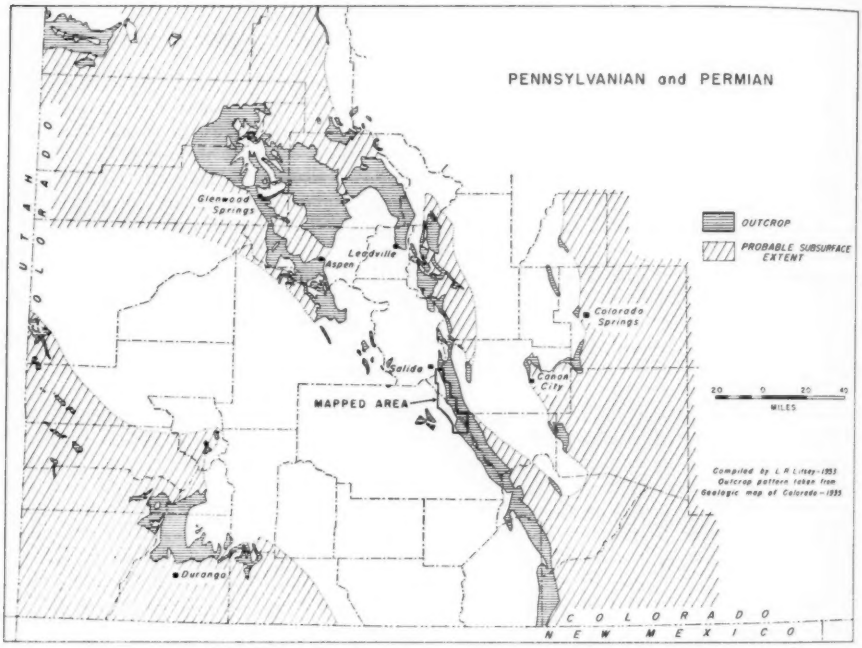

FIGURE 10.—EXTENT OF PENNSYLVANIAN AND PERMIAN SEDIMENTARY ROCKS

and a narrow band of upturned beds borders the eastern side of the Front Range and Wet Mountains.

The Pennsylvanian and Permian systems are discussed together here because accurate dating of several of the formations has not been possible, and definition of the boundary between the Pennsylvanian and Permian rocks is uncertain. The two periods represent a continuation of essentially the same geologic conditions. Sedimentary rocks of these periods record the only major Paleozoic uplift in Colorado. They are more than 14,500 feet thick in the northern Sangre de Cristo Mountains.

Positive areas in Colorado, known collectively as the Ancestral Rocky Mountains, consisted of two highlands: the Front Range highland, and the Uncompahgre highland (Fig. 11). These highlands were uplifted intermittently throughout Pennsylvanian and part of Permian time to furnish a great volume of clastic material. Maher (1953) outlined the Apishapa-

The area of the northern Sangre de Cristo Range was located between the Uncompahgre and Front Range highlands in a subsiding trough that trended north-northwest and extended from northeastern New Mexico into northwestern Colorado. The oldest Pennsylvanian rocks in this trough are fine-grained clastic rocks and limestones, but younger rocks contain a large amount of coarse material. The lower beds are characteristically gray or black, and the upper are typically red or brown. During Pennsylvanian time considerable marine deposition occurred in the middle of the basin; along the margins deltaic and coarse continental beds were deposited. In the north-central part of the trough an evaporite basin developed during Middle Pennsylvanian time, in which large amounts of gypsum and some salt were deposited (Brill, 1952, p. 820). A second evaporite basin developed on the southwest side of the Uncompahgre highland. Permian sedimen-

tation was characterized by continental deposition.

Numerous formation names for Pennsylvanian and Permian strata of Colorado have appeared in print, owing, in part, to difficulties in correlation (Tweto, 1949, Table 2, p. 187). Brill (1952, p. 812) proposed a simplified nomenclature for Pennsylvanian and Permian strata of the Colorado and New Mexico zeugogeosyncline. According to him, four widespread lithologic units are recognizable: (1) the equivalent Belden and Kerber formations, (2) the Clastic member of the Sandia formation, most of which is in New Mexico, (3) the equivalent Minturn and Madera formations and (4) the equivalent Maroon and Sangre de Cristo formations. The Schoolhouse tongue of the Weber sandstone and the State Bridge formation are local units confined to the northern end of the trough. In general, nomenclature of the strata in southwestern Colorado has been kept distinct from that in central Colorado.

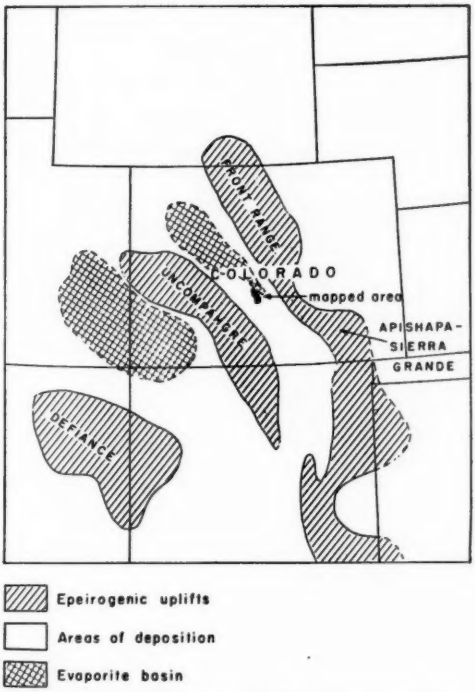

FIGURE 11.—PENNSYLVANIAN PALEOTECTONIC MAP
(Modified after Brill, Maher, McKee, and others.)

Kerber formation.—The Kerber formation was named by Burbank (1932, p. 13) for exposures along Kerber Creek in the Bonanza district west of Villa Grove, Colorado (Fig. 1). In its type section the Kerber consists of about 200 feet of white to gray coarse-grained sandstone and carbonaceous shale which overlie the Leadville limestone and extend upward to the base of the lowest red, micaceous sediment. Brill (1952, p. 814) suggested the name Kerber formation for the nonmarine facies of the Belden shale in the southern part of the trough.

The northern limit of beds that may be referred to the Kerber formation is uncertain; the southernmost outcrop of typical Kerber is east of Grayback Mountain in Huerfano Park (Burbank and Goddard, 1937, p. 941). In the vicinity of Wellsville, Brill (1952, p. 832) referred about 200 feet of dark-gray illitic shale, brown sandstone, and thin seams of coal to the Kerber formation. He noted the absence of the Kerber on Taylor Creek, 2 miles south of Wellsville. The only rocks similar to the Kerber found in the mapped area south of the

Bushnell Lakes are thin coaly shales and a few tens of feet of brown sandstone which separate these shales from the Leadville limestone in Lime Canyon. The total thickness of these beds probably does not exceed 150 feet.

No fossils other than plant remains have been found in the Kerber formation; hence, the age is unknown. Because of its stratigraphic position it may be Morrowan or Atokan.

Minturn formation.—The Minturn formation of Pennsylvanian and Permian (?) age was named by Tweto (1949, p. 194) for exposures along the east side of the Eagle Valley near Minturn, Colorado. Brill (1952, p. 819) extended the name Minturn to include equivalent beds throughout the Colorado zeugogeosyncline except in Moffat and Costilla counties, Colorado. In the type locality the Minturn is composed of lenticular beds of arkosic grit, shale, conglomerate, sandstone, quartzite, and a few relatively persistent beds of dolomite and limestone. Here, the contact of the basal part of the Minturn and the underlying Belden formation is gradational. The top of the Minturn is marked by the Jacque Mountain limestone member, which is the highest persistent limestone in the section.

The type section of the Minturn formation is located along the eastern margin of the basin of deposition, and the lithology changes toward the center of the basin. In many areas the Jacque Mountain limestone member is absent, and it is difficult to establish an adequate boundary between the Minturn and the overlying Maroon formation. A characteristic color change occurs throughout the basin west of the type section; the lower rocks are gray or brown and the upper beds are red. This color change is roughly parallel to, and somewhat above, the main zone of limestone in the section. Although this color change may neither correspond to a time line nor to the exact position of the Jacque Mountain limestone member, it has been used by some geologists to mark the top of the Minturn formation or its equivalents (Langenheim, 1952, p. 551).

Gypsum occurs sporadically in the Minturn formation from Meeker, Colorado, northwest of Glenwood Springs, southward to Coaldale; salt occurs locally in western South Park. Gypsum has been mined since 1902 from the Swissvale gypsum member (named by Brill, 1952, p. 833) near Coaldale.

The sedimentary rocks, mainly sandstones and conglomerates, assigned to the Minturn formation are approximately 8000 feet thick north of Rito Alto Peak as determined from aerial photographs and field measurements. This is approximately twice the thickness at Wellsville (Brill, 1952) and considerably greater than thicknesses reported farther south in Colorado. North of Rito Alto Peak the Minturn formation can be divided roughly into four main lithologic units, but these units apparently cannot be recognized very far from this locality. The units are defined for convenience in this report but are not proposed as formal stratigraphic units.

UNIT A: The basal unit is composed of approximately 400 feet of gray or brown, fine- to coarse-grained, quartzose sandstones interbedded with gray, micaceous, fine-grained sandstone or dusky-yellow sandstone.

UNIT B: Grayish red, fine- to coarse-grained sandstones interbedded with olive-gray, quartzose sandstones and fine conglomerates, up to 1200 feet thick.

UNIT C: The thickest unit comprises about 5700 feet of olive, greenish-gray, or brown sandstones, many of which are micaceous. There are thin clastic limestones, shales or siltstones, and conglomerates in minor amounts. The coarsest conglomerate observed contained pebbles 1 inch in diameter.

UNIT D: About 700 feet of interbedded gray limestones and brown sandstones. The thickness and composition of this unit is extremely variable.

According to D. W. Bolyard (oral communication), the entire Pennsylvanian and Permian sequence south of the mapped area can be roughly divided into three major units. The lowest unit is apparently equivalent to units C and D as used in this paper; it is uncertain if beds equivalent to units A and B are present. The middle unit of Bolyard is equivalent to the Sangre de Cristo formation of this report, and the upper unit is a younger conglomeratic sequence.

The Minturn near Rito Alto Peak is poorly stratified. In general, the sandstones are poorly sorted and composed of angular to subangular grains. Many of the sandstones are micaceous. There are few arkoses, but some sandstones are arkosic. In general the clastic rocks become coarser to the south.

The uppermost limestone of unit D is tentatively correlated with Brill's Whiskey Creek Pass limestone member of the Madera formation (1952, p. 832); it may be equivalent to the Jacque Mountain limestone member of the Minturn formation. According to Tweto (1949) the Jacque Mountain may be of Permian

age. The limestone in unit D is the only prominent limestone of the Pennsylvanian and Permian section in the southern part of the area described in this report. It ranges from 25 to 150 feet thick and is composed of massive- to thin-bedded clastic limestone. At one place it contains coarse quartz grains and is ripple-marked. The clastics above the limestone contain arkose, but those below do not.

According to Brill (1952, p. 820), the Minturn formation is Des Moines in age, except for the basal part, which may be of late Atoka age in some localities. It correlates with the Madera formation of northern New Mexico, the upper part of the Morgan formation of northwestern Colorado, and with most of the Hermosa formation of southwestern Colorado. The lower part of the Fountain formation along the eastern flank of the Front Range is also probably a correlative.

Sangre de Cristo formation.—The Sangre de Cristo formation, as defined by Hills (1899), comprised the entire Pennsylvanian and Permian section in the Sangre de Cristo Mountains. Brill (1952, p. 821) proposed that the term Sangre de Cristo formation be redefined to include Pennsylvanian and Permian strata overlying the Whiskey Creek Pass limestone member of the Madera formation or its equivalents.

The formation, as redefined, crops out from Pecos, New Mexico, to as far north as the Arkansas River near Salida. A geographic division between the Sangre de Cristo formation and the Maroon formation, its equivalent to the north, has been arbitrarily placed at the Fremont County-Park County line by Brill (1952, p. 822). The thickness ranges from 5500 feet near Crestone (Fig. 1) to more than 9600 feet near Whiskey Creek Pass, which is a few miles north of the southern boundary of Colorado. Southeast of Wellsville the thickness is 8800 feet (all thicknesses from Brill, 1952). South of Rito Alto Peak a thickness of about 6500 feet was measured by the author, but the top of the section was not reached.

Piedmont cyclothems, consisting of arkosic conglomerate grading upward to finer-grained material and, in places, to nodular limestone, characterize the formation. In the area investigated, the Sangre de Cristo formation is primarily conglomeratic arkose and fine-grained, red, micaceous sandstone but includes some interbedded gray sandstone. On the north side of Cotton Creek near the western margin of the sediments thick beds of conglomerate contain cobbles as much as 6 inches in diameter; also present are thin nodular limestone beds. To the south, the upper part of the Sangre de Cristo formation is the Crestone conglomerate of Melton (1925); east of Crestone it contains boulders more than 6 feet in diameter.

The exact age of the Sangre de Cristo formation is not known. Bones of pelycosaurs and of the cotylosaur *Didactes* sp. have been reported from a greenish-gray shale 1453 feet above the base of the formation in the Wellsville area (Brill, 1952, p. 834). Because these reptile bones have been reported from the Wolfcamp of Texas, Brill assigned the entire Sangre de Cristo formation to the Wolfcamp. Most writers consider the formation as Pennsylvanian and Permian.

Structure

Regional Structures

The Sangre de Cristo Range is the southernmost range in the second chain of the southern Rocky Mountains west of the Great Plains. The first chain, composed of the Front Range and the echelon Wet Mountains, is separated from the second chain by a series of basins called North, Middle, and South parks, and the combined Wet Mountain Valley and Huerfano Park (Fig. 1). Northward from the Sangre de Cristo Range the Mosquito, Gore, and Park ranges form a roughly linear topographic feature. Separating the Sangre de Cristo Range from the San Juan Mountains to the west is the broad, downfaulted San Luis Valley.

Compressive forces during late Cretaceous and early Tertiary (Laramide) time developed a series of long, meridional arches and basins. Many of these folds were broken by east-dipping thrust faults. West of the South Park basin a large anticlinorium developed, which now includes the Sawatch and Mosquito ranges and the intervening Arkansas River Valley and the northern Sangre de Cristo Range. The Sawatch Range is flanked on the west side by beds which have a westerly regional dip, and the Mosquito Range is composed of Paleozoic sedimentary rocks, which have a regional dip to the east. Gabelman (1952, p. 1594) referred to this anticlinorium as the Sawatch arch.

There is no known northern structural termination of the Sangre de Cristo Range, only a topographic boundary where the

Arkansas River swings eastward from Salida to Wellsville. The northern end of the range is composed of Precambrian rocks, which are part of the core of the Sawatch arch. The sedimentary rocks are part of the east flank of the Sawatch arch. They strike north to north-northwest and maintain an easterly dip, generally greater than 30°, from Wellsville south to the Orient mine. The contact between the Precambrian and Paleozoic rocks extends from the Mosquito Range across the Arkansas River into the Sangre de Cristo Range about 1 mile west of Wellsville (Fig. 1). To the south the contact is exposed progressively higher on the east side of the range and at Hayden Pass occurs on the crest. This contact becomes progressively lower southward on the west side of the range and near the Orient mine intersects the edge of the San Luis Valley and is covered by Quaternary gravels (Pl. 1). South of Hayden Pass the crest of the range is composed of Pennsylvanian and Permian strata, except where they have been intruded by three Tertiary stocks. The apparent reason for this outcrop pattern is the slight divergence between the axis of the Sangre de Cristo Range and the axis of the Sawatch arch.

The San Luis Valley and the modern Sangre de Cristo Range are of later origin than the Laramide features previously discussed. These younger large structural features are similar to the basin-and-range structures. One of the most prominent structures in this region is a concealed fault, probably steeply dipping, that separates the broad, flat San Luis Valley, which is partly filled with sediments of Cenozoic age, from the rugged mountains to the east. The San Luis Valley is apparently downthrown several thousand feet by this fault; the throw probably decreases northward from the Orient mine.

Local Structures

Introductory statement.—The part of the Sangre de Cristo Range from Wellsville to the Orient mine (Pl. 1) is characterized by a zone of high-angle strike faults near the base of the Paleozoic section. Many are reverse faults which dip steeply to the east. Several folds also affect the lower Paleozoic rocks; all have axes that strike north-northwest parallel to most of the faults. The folds are asymmetric with axial planes that dip steeply to the east.

Pennsylvanian and Permian beds are arched into a prominent diapiric anticline along the crest of the range south of the Rito Alto stock (Burbank and Goddard, 1937, p. 953). This stock intruded near the axis at the north end of the anticline was originally interpreted as the Precambrian core of the range (Endlich, 1877), which was then believed to be largely anticlinal.

South of Major Canyon, a large, west-dipping, high-angle, reverse fault brings Precambrian rocks on the west side into contact with Pennsylvanian and Permian rocks to the east. This fault is parallel to and about 1 mile east of the foot of the range and extends southward beyond the limit of the mapped area.

Detailed structural features of the northern Sangre de Cristo Range can be conveniently grouped into seven areas, with closely related structures. From north to south these areas are: (1) Wellsville, (2) Calcite, (3) Bushnell Lakes, (4) Hayden Pass, (5) Steel Canyon and Orient, (6) Rito Alto Peak, and (7) Major Canyon (Fig. 12). Several of the more important faults and folds have been named for the most prominent topographic feature in the immediate vicinity. Block diagrams of parts of four of these areas are presented (Fig. 12).

Wellsville area.—The following summary is taken from J. W. Rold (1950, unpub. MS thesis, Univ. Colorado).

Structures in the Wellsville area can be divided into three types; folds that trend northwest, reverse faults that trend northwest and dip 40°–60° northeast or southwest, and vertical faults that trend west-northwest. Most of the folds are asymmetric. Each fold is closely associated with and parallel to one or more reverse faults. All folds plunge southeastward and apparently die out without markedly affecting the Pennsylvanian beds.

The Lofgren anticline, the northernmost fold, is truncated to the northwest by an intrusive body (Pl. 1). It is cut on its southwest flank by a parallel reverse fault, which dips about 45° northeast under the fold. The fault has a maximum displacement of about 600 feet, but it could not be traced more than a quarter of a mile in either direction from the point of maximum displacement.

Along its southeast projection the Lofgren anticline is replaced by the Wells Gulch syncline which is overridden by the upthrust blocks of two parallel reverse faults of opposing dip. One of these overrides the northeast limb of the fold. This fault dips approximately 40° northeast, but where it cuts the less compe-

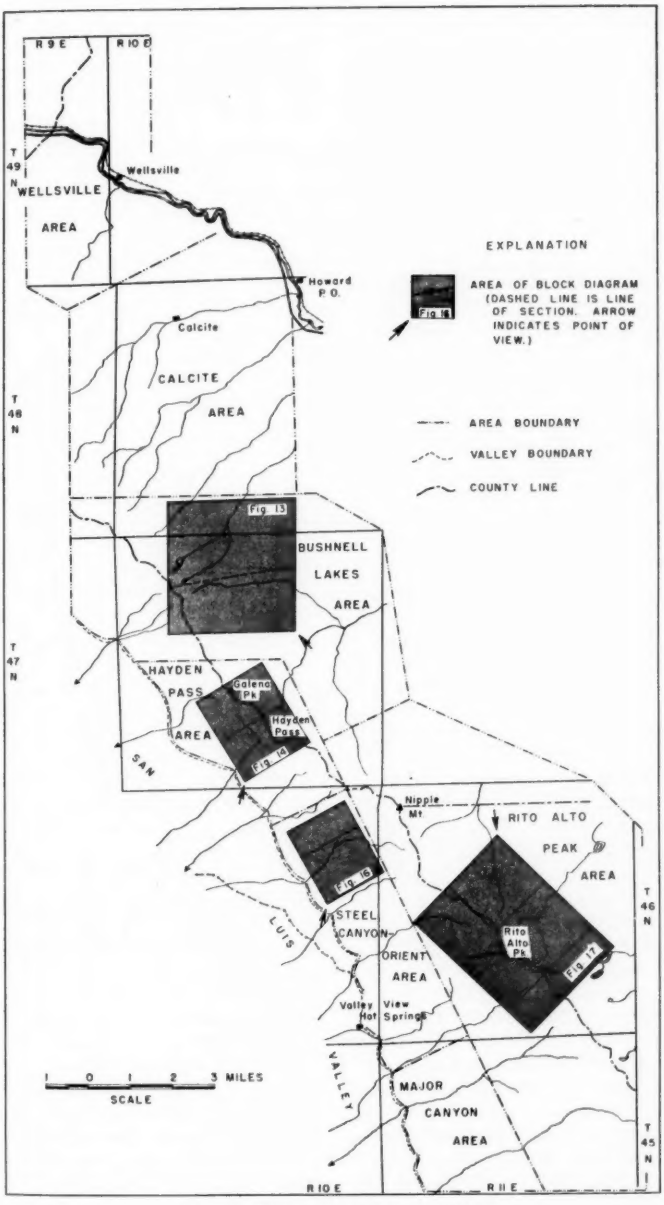

FIGURE 12.—STRUCTURE INDEX MAP

tent Pennsylvanian rocks, the dip steepens to about 60°, and the fault breaks into two or possibly more smaller branches. The only west-dipping reverse fault in this vicinity overrides the syncline from the southwest. The fault dips about 50° southwest; the stratigraphic displacement is about 175 feet. This fault, like the one on the northeast flank of

the Wells Gulch syncline, seems to steepen appreciably and to break into smaller faults where it transects Pennsylvanian rocks.

The largest fold in the area is the Wellsville syncline, which is well exposed in the bend of the Arkansas River north of Wellsville. It plunges southeastward and seems to die out in that direction, as it is not manifested in the Pennsylvanian rocks southeast of Wellsville.

The only other important fold in the Wellsville area is the Rogers anticline, which parallels the Rogers fault and probably resulted from the same stresses. The intensity of folding decreases to the southeast, whereas the fault displacement increases in that direction.

A series of vertical faults (Box Canyon, Quarry, Norberg, and two smaller faults) trend approximately N. 65° W. The north side of each fault is upthrown, and the beds north of each fault are offset eastward. Vertical Devonian and Mississippian beds are offset about a quarter mile on the east side of Taylor Gulch by the Norberg fault, suggesting horizontal movement of the southwest block to the northwest. Similar movement along the other vertical faults is suggested by drag in the plan view, especially where the beds dip steeply and the faults are nearly perpendicular to the strike of the beds.

Calcite area.—The following summary is taken from W. D. Siapno (1953, unpub. MS. Thesis, Univ. Colorado).

From Norberg fault southward for about 4 miles the Pennsylvanian rocks are in contact with the Mississippian rocks along the Travertine fault, a fault which strikes a few degrees west of north and dips steeply to the east. Exact displacement could not be calculated. The Travertine fault has been displaced by several northeast-trending vertical faults.

At two places in the Calcite area, high-angle, reverse strike faults repeat all or part of the pre-Pennsylvanian sequence. Two of these faults are west of Calcite; a third is south of West Creek in section 20, T. 48 N., R. 10 E. These reverse faults seemingly terminate abruptly against the northeast-trending vertical faults or against the northwest-trending vertical Norberg fault.

One of the most perplexing structures in the calcite area is the boundary of an outlier of sedimentary rocks of Paleozoic age surrounded by Precambrian rocks and located about 1½ miles southwest of Calcite. All formations from Ordovician to Pennsylvanian are present in normal depositional sequence and dip northeast in the outlier. The limits of the outlier are poorly defined, but it is inferred that the block of sedimentary rocks is entirely surrounded by a fault or faults.

Bushnell Lakes area.—High on the east side of the range north of Galena Peak, pre-Pennsylvanian sedimentary rocks are folded into the asymmetrical Bushnell syncline, which is bounded by Precambrian metamorphic rocks both to the east and west (Pl. 1). The axial plane dips 60°–70° E.; the axis is apparently horizontal and ranges in strike from due north to N. 18° W. Beds on the west flank on the syncline dip about 30° E., but beds on the east flank range in dip from 59° W. to slightly overturned, increasing from north to south.

The Bushnell syncline is bounded on the east by the Bushnell fault, a high-angle reverse strike fault that brings Precambrian rocks into contact with the Ordovician sequence (Fig. 13). The trace of the fault parallels the trace of the synclinal axis; both are convex toward the west, and it is inferred that both structures were produced by the same forces. The steep dip of the east limb of the syncline was probably produced by drag along this fault. The fault plane is nearly vertical at the surface, but because of the asymmetry of the Bushnell syncline, the dip is inferred to be steep to the east with increasing depth. Movement on the fault probably dies out to the north where the dip of the beds west of the fault becomes lower. Maximum displacement could not be determined.

The steeply-dipping Hayden Creek fault (Gabelman, 1952, p. 1606) is about 1 mile east of the Bushnell fault. The east side is downthrown, in most places bringing Pennsylvanian beds on the east side into contact with Precambrian rocks on the west. The fault strikes north-northwest along its southern exposure, but swings to the northwest as it is traced northward. W. D. Siapno (1953, unpub. MS thesis, Univ. Colorado) followed this fault northward for about 2 miles; there it is interrupted or terminated by a transverse fault. Hayden Creek fault has not been observed south of the middle prong of Hayden Creek, but it may continue some distance southward in the Pennsylvanian rocks.

The exact displacement along the Hayden Creek fault could not be calculated, but certain approximations can be made. Leadville limestone is exposed on the east side of the fault opposite Precambrian rocks near the southern

end of the fault (Pl. 1). About a mile to the north the Manitou formation crops out on the west side of the fault opposite Pennsylvanian rocks. Near the outcrop of the Manitou the fault has a minimum throw of about 1100 feet and a maximum throw of probably less than 2500 feet.

Although covered by glacial debris, a fault striking about N. 60° W. is inferred at the southern end of the horst of Precambrian rocks, because unfolded, east-dipping lower Paleozoic rocks are present south of the covered area along the trend of the Bushnell fault and syncline. This covered fault probably developed contemporaneously with the Bushnell fault and Bushnell syncline to allow essentially vertical movement of the north block upward past the undisturbed south block.

Three small reverse faults are present along the west flank of the Bushnell syncline, each with the east side upthrown, and having but a few tens of feet of displacement. These faults have the same type of movement as the Bushnell fault and were probably produced by the same forces.

A syncline in the Pennsylvanian rocks in the eastern part of this area has been interpreted from field data and aerial photographs. Exposures are poor, and bedding is difficult to recognize, making interpretation difficult. The syncline seems to strike north-northwest.

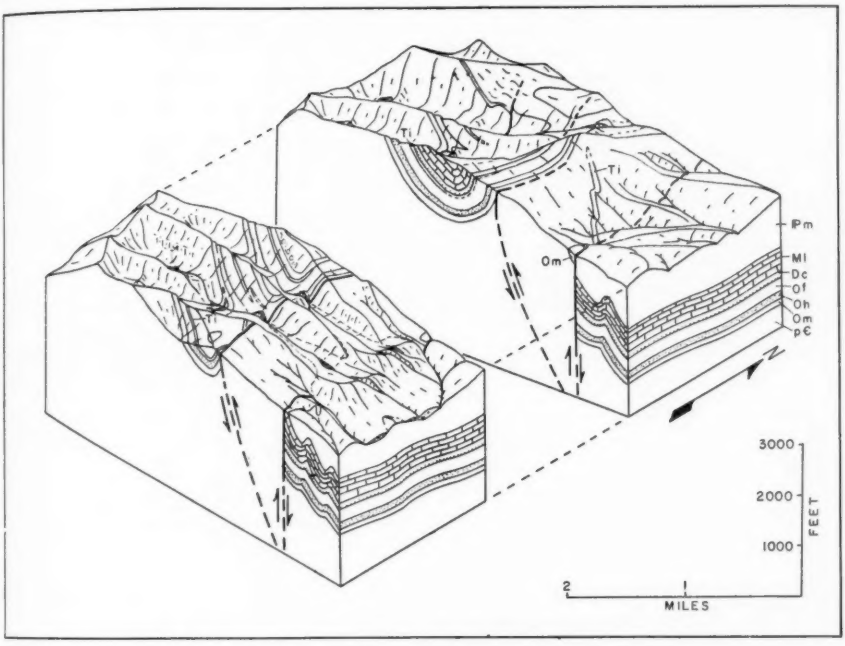

FIGURE 13.—BLOCK DIAGRAM OF BUSHNELL LAKES AREA

Although the individual folds could not be mapped, a number of small, tightly folded anticlines and synclines must exist in the Pennsylvanian rocks east of Hayden Pass and east of the southern end of the Hayden Creek fault to explain the numerous reversals of dip noted in the field. These minor folds all seem to have a northerly trend, generally parallel to the major structural features.

Hayden Pass area.—The sedimentary rocks of Paleozoic age cross the crest of the range on the south side of Galena Peak in a small asymmetrical anticline that plunges southeast (Fig. 14). To the west the beds pass into an asymmetrical syncline. Both folds are broken near their axes by high-angle reverse faults that strike northwest, parallel to the fold axes, and dip steeply northeast.

The Galena Peak fault, which passes just west of Galena Peak, appears to be a hinge fault that begins near Hayden Pass and in-

creases in throw to the northwest. A tongue of the Manitou formation extends about 1800 feet northwestward from Galena Peak on the west side of the fault; beyond this, the anticline and a faulted syncline to the west of it. These folds extend from a few hundred yards north of Steel Canyon southward more than 4 miles to the Orient mine. The Lead-

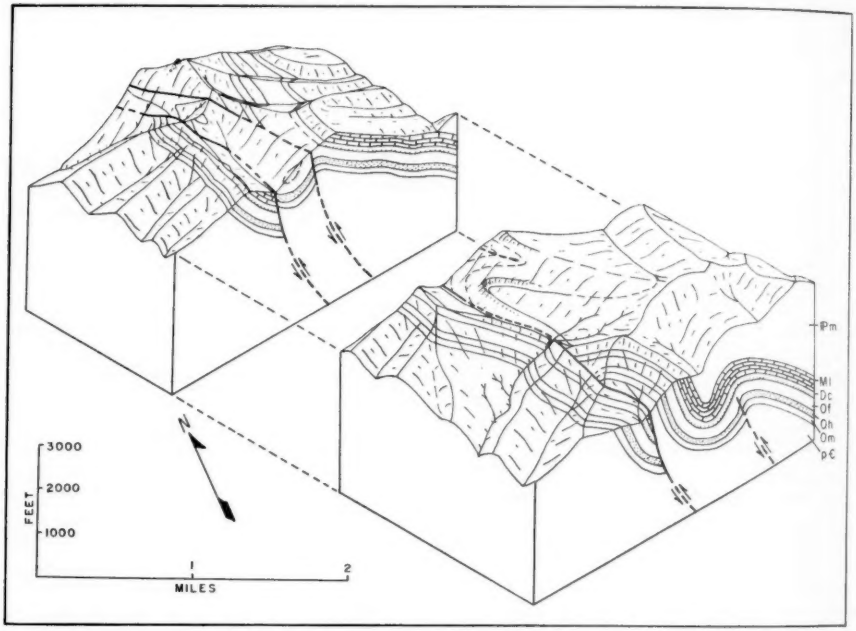

FIGURE 14.—BLOCK DIAGRAM OF HAYDEN PASS AREA

fault is evident for a short distance in the Precambrian metamorphic rocks.

The Silver Creek fault is west of the synclinal axis and parallel to it. Along the ridge southwest of Galena Peak this fault brings dolomite of the Fremont formation on the east against Leadville limestone on the west; at this point the throw is approximately 500 feet. This fault extends an unknown distance into Precambrian metamorphic rocks to the northwest. Displacement along this fault increases to the south, where the Silver Creek fault brings Leadville limestone on the west into contact, successively, with Mississippian, Devonian, Ordovician, and Precambrian rocks. The fault extends into Precambrian rocks south of Silver Creek, but its full extent is unknown. Where the Harding sandstone is in contact with the fault on the east side, curved bedding planes appear to be the result of drag, suggesting that the east side moved northward as well as up.

Steel Canyon-Orient area.—The tightest folds in the area of this report are the Steel Canyon anticline and a faulted syncline to the west of it. These folds extend from a few hundred yards north of Steel Canyon southward more than 4 miles to the Orient mine. The Leadville limestone forms cliffs and ridges in this region and produces striking topographic expression of the folds (Figs. 15 and 16). Beds older than Mississippian are exposed in each canyon along the anticlinal axis. The east limb of the anticline dips generally 50°–60° northeast; the steeper west limb is vertical in some places.

Pennsylvanian beds occupy the axial region of the syncline nearly as far south as Lime Canyon, and a small outcrop of Pennsylvanian rocks on the next ridge to the south indicates the position of the axis. South of Black Canyon the syncline is more open and plunges to the south. The pre-Pennsylvanian rocks on the west limb of the syncline swing to the southwest and are truncated by the Sangre de Cristo fault at the Orient mine. South of the Orient mine Pennsylvanian beds are adjacent to the fault.

The west flank of the syncline is cut by three nearly vertical strike faults. Locations of the faults could not be determined exactly, but they are limited by prominent ridges of Lead-

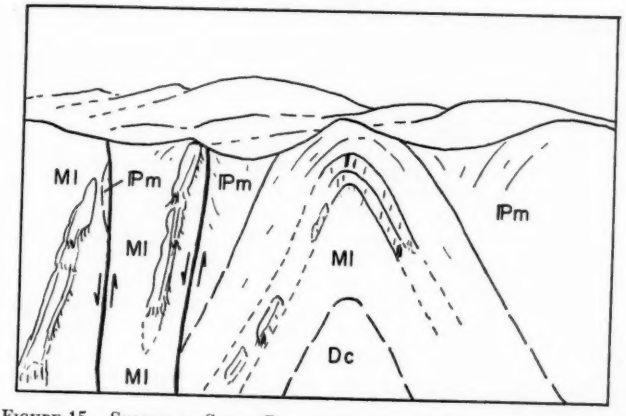

FIGURE 15.—SKETCH OF STEEL CANYON ANTICLINE AND FAULTED SYNCLINE
Viewed looking northwest across Lime Canyon. (Dc—Chaffee formation, Ml—Leadville limestone, Pm—Minturn formation.)

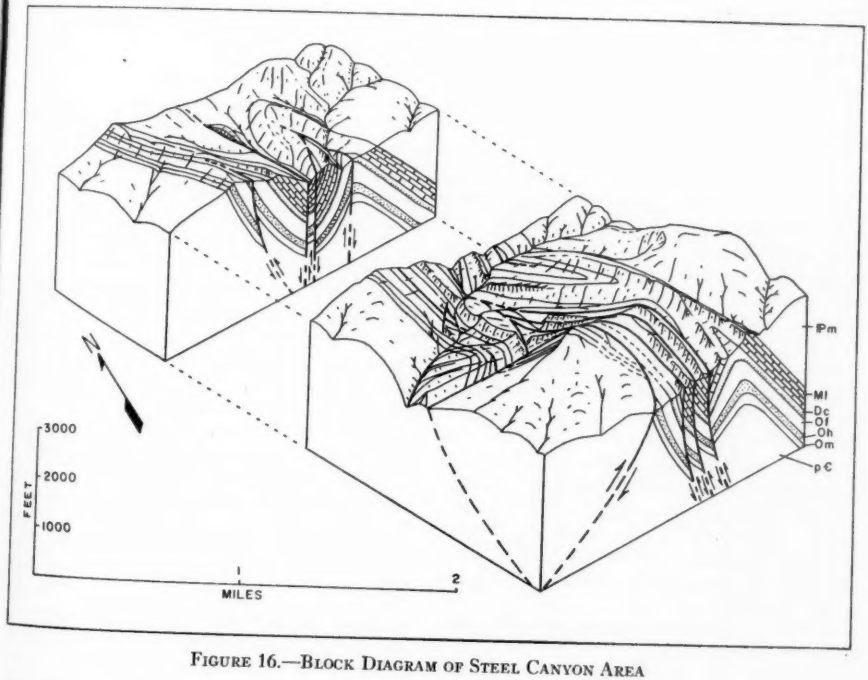

FIGURE 16.—BLOCK DIAGRAM OF STEEL CANYON AREA

ville limestone on the north side of Lime Canyon, where the formation dips steeply. The east side of each fault is relatively upthrown; displacement along any one fault is probably less than 1000 feet.

To the north these three strike faults terminate against or join a fault that strikes nearly north (Pl. 1). The pre-Pennsylvanian beds on the west side of this fault are continuous northward but change in strike from north to north-northwest.

These pre-Pennsylvanian rocks are terminated to the south by the westernmost fault in the Steel Canyon-Orient area. This fault

has an irregular trace and extends from Steel Canyon to Black Canyon. Although this fault could not be traced continuously because of poor exposures, it is interpreted as a reverse fault that strikes northwest and dips about 45° southwest. Near Steel Canyon steeply dipping Ordovician rocks on the southwest side of the fault are nearly perpendicular to beds north of the fault. On the ridge between Lime Canyon and Black Canyon, pre-Pennsylvanian beds southwest of the fault are displaced eastward several hundred feet.

South of Black Canyon, the Steel Canyon anticline is broken by a series of parallel, minor strike faults, inferred from repetition of Mississippian and Pennsylvanian beds. Inferred structural relationships are shown on Plate 1, section *D-D'*. These faults do not appear to be continuous with the strike faults north of Black Canyon.

Gabelman (1952, p. 1592) suggested a fault in Black Canyon to explain offsetting of contacts across Black Canyon. In addition, fold axes and strike faults cannot be projected across the canyon. The fault, however, is covered throughout its length by glacial deposits. The outcrop of Pennsylvanian rocks in the center of the syncline is wider south of the canyon, suggesting that the south side is relatively downthrown.

The boundary between the Sangre de Cristo Mountains and the San Luis Valley is a sharply defined, linear feature in the area of this report and to the south. This boundary is interpreted as a large normal fault (*see* Burbank and Goddard, 1937, p. 974) and is referred to in this report as the Sangre de Cristo fault. The trace of the fault is concealed by Quaternary gravels, and the fault plane has not been observed, but its presence is suggested by the following features: (1) the linear, well-defined trend of the mountain and valley boundary. The ends of the mountain ridges have the appearance of eroded triangular facets, although no positive evidence was found to support this contention; (2) truncated structures at the Orient mine and at Major Canyon; and (3) the presence of hot springs at the edge of the valley at Valley View Hot Springs (Pl. 1). The difference in elevation between the pre-Tertiary floor of the San Luis Valley and the highest peaks in the range is unknown, but it probably exceeds 10,000 feet. From this it is inferred that the San Luis Valley is downthrown several thousand feet.

Brecciated and silicified sedimentary rocks, believed to be the Harding sandstone and Fremont formation, are exposed between the Sangre de Cristo fault and sandstones of Pennsylvanian age near Valley View Hot Springs. These sedimentary rocks appear to be repeated by one strike fault and separated from the Pennsylvanian rocks by another. These faults may represent the northern extension of a prominent zone of reverse faults described in the section on the Major Canyon area.

Pennsylvanian sedimentary rocks have been metamorphosed into phyllites and other low-rank metamorphic rocks in the area immediately north of Garner Canyon and adjacent to the Sangre de Cristo fault. The degree of metamorphism decreases gradually to the east, and sedimentary rocks to the north and south are not strongly metamorphosed. Foliation is apparently parallel to bedding, and both strike nearly parallel to the Sangre de Cristo fault at this point.

Rito Alto Peak area.—The Pennsylvanian sedimentary rocks in the vicinity of Rito Alto Peak have been intruded by the Rito Alto stock which is exposed over about 4 square miles. The intrusion of the stock did not greatly modify pre-existing structures, although beds adjacent to the contact are vertical or overturned in a few places (Fig. 17). P. Toulmin III (1953, unpub. MS Thesis, Univ. Colorado) stated that structures in the country rock affected by intrusion were noted only in the vicinity of the bodies of intrusive rhyolite. From this he suggested that the rhyolite may have been the only phase of the stock that was forcefully intruded.

In general, structure along the western border of the stock is obscure. At the head of Black Canyon, beds near the stock dip steeply and strike north-northwest, nearly parallel to the igneous contact (Pl. 1). Farther south, at the head of Garner Canyon, beds strike about N. 40° E., nearly perpendicular to both the stock contact and to the strike of the sediments a short distance to the north; the dip ranges from 55° to 70° southeast. This trend, which is transverse to the major structural elements both to the north and south, appears to be continuous westward to the foot of the range. The reasons for this transverse structural trend are unknown.

Pennsylvanian rocks are folded into the Nipple Mountain syncline along the crest of the range north of the stock where Saguache, Fremont, and Custer counties join. The syn-

clinal axis strikes approximately N. 25° W.; its plunge could not be determined. The west flank of this syncline is well exposed along the crest of the range. The east flank is difficult to define, but a few confirmatory attitudes were obtained east of Nipple Mountain. An anticline approximately 1 mile east of Nipple Mountain has been interpreted from aerial photographs.

Burbank and Goddard (1937, p. 954) describe the anticline south of Rito Alto stock as a diapiric fold. The eastern flank is well exposed east and southeast of the stock; the beds have a uniform strike of about N. 25° W. and dip northeast. This dip steepens sharply from about 40° to 60° and, in places, becomes vertical as the anticlinal axis is approached. Structure on the western limb is obscure; beds strike northwest and dip 30°–50° southwest. The fold extends several miles to the south, and the axis gradually crosses from the west to the east side of the crest of the range in that direction.

Major Canyon area.—The faulted area near Major Canyon is continuous with a large fault zone that extends southward beyond Crestone (Fig. 1). This zone is a composite, west-dipping, high-angle reverse fault that is postulated by Burbank and Goddard (1937, p. 970) to connect with a thrust fault at Kerber Creek, on the west side of the San Luis Valley. The fault zone swings westward at Major Canyon and is truncated by the later Sangre de Cristo fault.

Precambrian rocks are exposed along the foot of the range west of the fault zone; Pennsylvanian or Permian rocks lie east of it. Isolated segments of pre-Pennsylvanian strata are present in the fault zone. Individual blocks within the fault zone are separated by high-angle faults, probably reverse faults, that

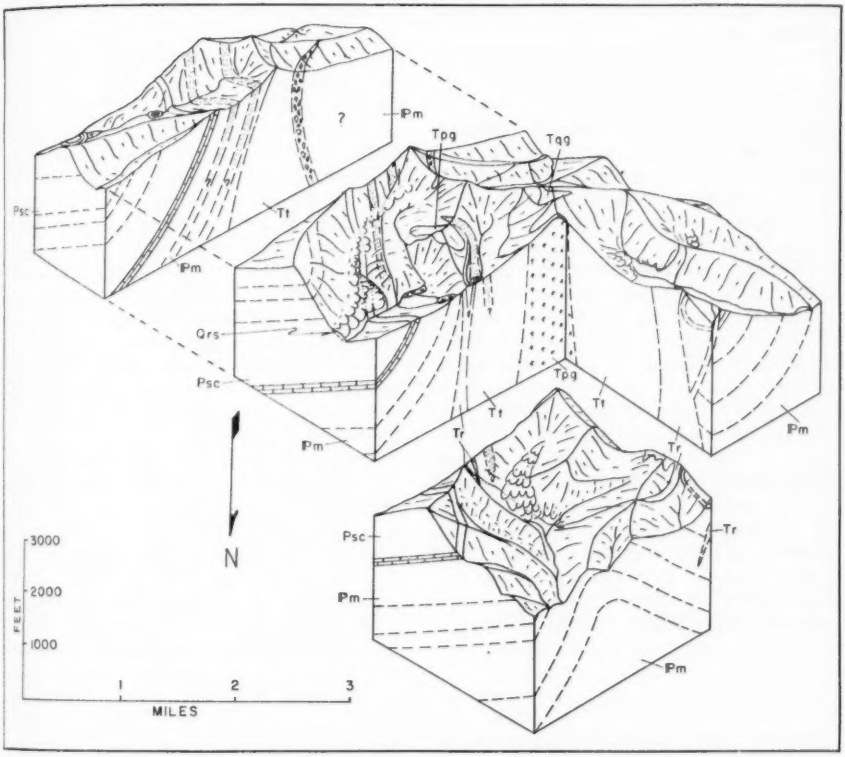

FIGURE 17.—BLOCK DIAGRAM OF RITO ALTO PEAK AREA

strike north, and by transverse or tear faults that strike east.

The complex fault zone is at least half a mile wide, and individual segments of pre-Pennsylvanian strata give ample evidence of structural complexity within this zone. On one ridge in the eastern half of section 18, T. 45 N., R. 11 E., a partial section is repeated four times within 400 yards. Ridges to the north and south show an entirely different sequence. In addition, the Manitou formation and Harding sandstone are almost vertical near the Precambrian contact to the west in the eastern halves of sections 7 and 18 but to the east they are overturned to a nearly horizontal position.

A complete section of brecciated, pre-Pennsylvanian sedimentary rocks crops out on the north side of Major Canyon. The beds strike nearly due east and are overturned, dipping about 50° south. These beds are at the western end of the zone of transverse structures that begins at the western margin of the Rito Alto stock as discussed in the preceding section. Three parallel, north-striking faults displace the formation contacts from 100 to 300 feet; the east side of each fault is relatively displaced to the north. Apparently an east-west fault follows Major Canyon, because Pennsylvanian rocks are exposed south of the canyon opposite the overturned earlier Paleozoic section. The fault may be part of the main thrust fault.

Tectonics

Laramide and later movements have produced structures in the northern Sangre de Cristo Mountains indicative of both horizontal and vertical forces. Strong compressive forces are evidenced by tight folds and reverse faults in the northern Sangre de Cristo Range and by prominent thrust faults in adjacent regions to the north and south. Vertical forces are indicated by the present high elevation of the Sangre de Cristo Range in comparison to the elevation of the downfaulted San Luis Valley. The pronounced parallelism of the faults and fold axes is the most prominent feature of the structure in the range. Most of the structures strike north-northwest.

Most of the structures within the northern Sangre de Cristo Mountains are apparently the result of compressive forces directed nearly east and west normal to the strike of the strike faults and fold axes. The strike faults from the Orient mine to Wellsville, with the possible exception of the Travertine and Hayden Creek faults, appear to be high-angle reverse faults. Most of these faults dip steeply northeast and have the northeast side upthrown. Some of these faults, however, are vertical, but because they have the northeast side upthrown and are closely associated with tight asymmetrical folds that have northeast-dipping axial planes, these vertical faults are also interpreted to result from compression. The prominent reverse fault that extends southward from Major Canyon strikes north-northwest, nearly parallel to the faults to the north, but it dips steeply to the west. Probably in all cases the vertical component of movement on the strike faults was greater than the horizontal component. Most of the transverse faults are also interpreted to result from compression and form a system of conjugate shears. All the folds mapped trend north-northwest parallel to the strike faults and apparently were formed at the same time and by the same forces that produced the reverse faults.

The Travertine fault and the Hayden Creek fault may be either normal or reverse faults; the evidence is inconclusive. The Travertine fault dips steeply northeast parallel to the reverse faults in the vicinity, and it is offset by small faults that strike northeast and are thought to be tear faults. The age relationship of the beds across the fault suggests that it is a normal fault; however, the fault is probably a reverse fault with younger beds (Pennsylvanian) thrust over older beds (Mississippian). The Hayden Creek fault may be a reverse fault with the west side upthrust, but by negative reasoning it is interpreted to be a normal fault. If the transverse fault between Cherry Creek and West Creek (Pl. 1) were a tear fault, and if the strike faults to the north and south were reverse faults, the relative movement along the transverse fault would be opposite to that observed. The rocks on the south side of the fault would be thrust eastward relative to the beds on the north side. Therefore the transverse fault is probably not merely a tear fault, but is a fault that has the south side downthrown. It may be an older tear fault that has also had later vertical movement. The Hayden Creek fault is probably a normal fault with the east side downthrown.

The reverse faults are largely confined to the boundary between the Precambrian and lower Paleozoic rocks, and the boundary between the Mississippian and Pennsylvanian

rocks. These are boundaries between rocks of different character and are also unconformities. The reverse faults are probably bedding-plane thrusts at depth that follow these two zones of weakness, and that increase in dip as they approach the surface, becoming high-angle reverse faults.

The largest fault in the area is the concealed Sangre de Cristo fault, which separates the Sangre de Cristo Range from the San Luis Valley. Burbank and Goddard (1937, p. 965) postulate that the fault was caused primarily by collapse and foundering of the San Luis Valley as a result of withdrawal of large volumes of magma from beneath the valley to produce the lava flows of the San Juan Mountains. Gabelman (1952, p. 1610) suggests that the whole area was uplifted and the San Luis Valley later downdropped by a basin-and-range type of faulting. The height and ruggedness of the modern Sangre de Cristo Range suggest that uplift of the range was an important factor. Uplift along the Sangre de Cristo fault probably produced eastward tilting of the range. This tilting would have steepened the older eastward-dipping faults.

Determination of relative ages of Laramide structures in the northern Sangre de Cristo Range is difficult and in some cases impossible. The major fold in the region and probably the oldest Laramide structure is the Sawatch arch. The smaller folds and high-angle reverse faults were probably formed later when compressive forces reached a climax. If the Hayden Creek fault is a normal fault, it is probably younger than the reverse faults. The transverse faults may vary considerably in age. Most of them are probably tear faults that formed simultaneously with the reverse faults. Others are probably normal faults and may be either older or younger than the reverse faults. The Sangre de Cristo fault is of later origin than the reverse faults.

Structural events other than those of the Precambrian cannot be dated more precisely than as post-Permian and pre-Pleistocene from data in the mapped area, because the youngest deformed rocks are Permian sedimentary rocks, and the oldest beds not involved are Pleistocene gravels. The age of the structures must be inferred from study of adjacent areas where Late Cretaceous and Tertiary rocks are involved. The nearest places where such situations exist are Huerfano Park, South Park, and the San Luis Valley.

In Huerfano Park, Upper Cretaceous beds are overlain by the Tertiary Poison Canyon, Cuchara, and Huerfano formations in that order. Burbank and Goddard (1937, p. 957–961) have interpreted the Laramide and post-Laramide structural history as follows:

(1) Uplift and erosion of the old Paleozoic landmass to the west took place in Paleocene time, as evidenced by the unconformity between the Vermejo formation (Late Cretaceous) and the Poison Canyon formation (Paleocene) and the presence of granitic material in the Poison Canyon formation.

(2) Deformation by overthrusting reached a climax between deposition of the Poison Canyon and Cuchara formations (early Eocene) but continued into Huerfano time (early and middle Eocene). This is indicated by the prominent angular unconformity beneath the Cuchara formation and the lesser unconformity beneath the Huerfano formation.

(3) Upthrusting of the Precambrian block along the west side of the present range in middle Eocene time.

(4) Downfaulting of the San Luis Valley between the late Oligocene and the end of the Tertiary.

In South Park, the east-dipping Elkhorn thrust fault cuts the Denver formation, which is considered to be Late Cretaceous and Paleocene in age (Stark *et al.*, 1949, p. 122); therefore, the South Park thrusting is post-Denver and probably post-Paleocene. There is nothing in South Park that contradicts the evidence in Huerfano Park.

The Sante Fe group of middle (?) Miocene to Pleistocene (?) age contains the oldest known Tertiary sedimentary rocks in the San Luis Valley (Siebenthal, 1910, p. 39). As the valley contains erosion products from the Sangre de Cristo Range and the San Juan Mountains after initial development of the valley, the Sangre de Cristo fault probably began no later than Miocene time. The Hinsdale formation of Pliocene (?) age, which slopes from the San Juan Mountains into the San Luis Valley (Atwood and Mather, 1924, p. 121), and the faulted Los Pinos gravel of Miocene and Pliocene (?) age near Fort Garland (Atwood and Mather, 1932, p. 99) indicate continued disturbance in the San Luis Valley. The ruggedness of the range suggests that present elevations are due to movements no older than Pliocene.

The meager evidence concerning the age of the Rito Alto stock does not permit accurate dating of the intrusion. The stock truncates Laramide structures and was eroded by Pleistocene glaciers. Because volcanic debris

is found in Huerfano Park in the Cuchara formation of early Eocene age, Burbank and Goddard (1937, p. 974–975) have dated the stocks of the Sangre de Cristo Range as early to middle Eocene.

Conclusions concerning the Laramide and later structural history of the northern Sangre de Cristo Mountains do not differ greatly from the interpretations of Gabelman (1952, p. 1609–1611): deformation began in Paleocene time with the development of the Front Range, Wet Mountains, and Sawatch arch, followed by tight folding and thrusting in the early Eocene. The Rito Alto stock was intruded probably in early to middle Eocene time. Prior to the beginning of Pliocene time the mountains began to be uplifted along the Sangre de Cristo fault. Other vertical faulting may have accompanied this uplift. Movements probably continued intermittently throughout the Pliocene and part of Pleistocene time. Late epeirogenic uplifts have affected the entire region.

Summary of Geologic History

The Precambrian history of the northern Sangre de Cristo Mountains has not been studied. During early and middle Paleozoic time this area was near the southern edge of a trough that extended from northwestern Colorado to southeastern Colorado between the positive areas of Siouxia and Sierra Grande. The area was intermittently inundated by seas in which blanket-type deposits were laid down. Periodic minor uplifts and erosion produced disconformities between all formations.

The Sangre de Cristo area probably remained positive throughout the Cambrian period, although the southern shore line of the Late Cambrian sea was only a few miles to the north. The Early Ordovician sea spread over this area, which had low relief; the Manitou formation was deposited, and the sea retreated. Invasions followed in Middle and Late Ordovician times, resulting in deposition of the Harding sandstone and the Fremont formation. During Silurian and Early Devonian time this area was slightly above sea level and was undergoing mild erosion. The Late Devonian sea spread eastward from the Cordilleran geosyncline, inundating this part of Colorado and depositing the Chaffee formation. As the sea advanced, reworked weathered material and small amounts of detritus added from the land areas were laid down as basal clastics. The overlying limestones and dolomites indicate deposition in quiet waters. In Early Mississippian time (Kinderhook and Osage), seas encroached upon Colorado from both east and west and met in central Colorado probably in the Late Mississippian (Meramec); near the end of this epoch the seas retreated eastward and westward. It has not yet been determined if all three of these epochs are represented by the Leadville limestone in the northern Sangre de Cristo Range. Widespread erosion in the Late Mississippian and Early Pennsylvanian removed unknown but unequal amounts of Mississippian strata.

In early Pennsylvanian time a trough, extending from northwestern Colorado through the Sangre de Cristo region into north-central New Mexico, began to sink slowly. The Morrow sea entered this trough from the northwest and extended to a point near Wellsville. Later, an Atoka sea from the south probably invaded the Sangre de Cristo area (Brill, 1952, p. 841). Probably a large part of the Pennsylvanian rocks in the Sangre de Cristo area were deposited in the Des Moines sea, which was more extensive in Colorado than other Pennsylvanian or Permian seas. Intermittent uplift of the Ancestral Rocky Mountains in Colorado resulted in deposition of interbedded coarse continental sediments and thin marine beds. Numerous evaporite basins developed during Des Moines time in central Colorado as far south as Coaldale. Seemingly, an ecologic barrier formed across the trough in the area beginning near Hayden Pass and extending south beyond the area described in this report. There are lithologic differences between the sedimentary rocks in the area of the barrier and the areas to the north and south. In Missouri and Virgil times the trough was probably uplifted above sea level.

The greatest uplift of the Uncompahgre highland west and southwest of the northern Sangre de Cristo Range occurred in Late Pennsylvanian or Early Permian time, but the highlands were probably low by the end of the Permian period. During the Permian period the central part of the trough, including the Sangre de Cristo area, remained positive and received continental sediments, but the northern and southern ends were invaded by shallow seas. The area south of Hayden Pass was in the most rapidly sinking part of the trough and received more than 14,500 feet of Pennsylvanian and Permian clastic sediments.

During the Mesozoic, sediments were undoubtedly deposited in the northern Sangre

de Cristo Mountains, but they were removed by erosion during and following Laramide uplifts.

In the northern Sangre de Cristo Range, which is on the east flank of the Sawatch arch, compressive forces during the Laramide revolution produced a series of tight folds and high-angle reverse faults with a predominant north-northwest trend. The Rito Alto stock was intruded during the latter part of the Laramide orogeny, probably in Eocene time. Intrusion was accompanied by hydrothermal activity and some mineralization.

In middle Tertiary time prior to the beginning of the Pliocene the present Sangre de Cristo Range was uplifted and tilted eastward along the Sangre de Cristo fault, which bounds the range on the west side. Intermittent movements continued through the Pliocene and into the Pleistocene.

Alpine glaciers deeply scoured the higher parts of the range during at least one stage of the late Pleistocene glaciation.

References Cited

Atwood, W. W., and Mather, K. F., 1924, Physiographic history of the the San Luis Valley of Colorado and New Mexico (Abstract): Geol. Soc. America Bull., v. 35, p. 121–123

—— 1932, Physiography and Quaternary geology of the San Juan Mountains, Colorado: U. S. Geol. Survey Prof. Paper 166, 176 p.

Bagg, R. M., Jr., 1908, Some copper deposits in the Sangre de Cristo Range, Colorado: Econ. Geology, v. 3, p. 739-749

Bass, N. W., 1944, Paleozoic stratigraphy as revealed by deep wells in parts of southwestern Colorado, northwestern New Mexico, northeastern Arizona, and southeastern Utah: U. S. Geol. Survey Prelim. Chart 7, Oil and Gas Investig. Ser.

Bass, N. W., and Northrop, S. A., 1953, Dotsero and Manitou formations, White River Plateau, Colorado, with special reference to Clinetop algal limestone member of Dotsero formation: Am. Assoc. Petroleum Geologists Bull., v. 37 p. 889–912

Bassett, C. F., 1939, Paleozoic formations of Dotsero Colorado: Geol. Soc. America Bull., v. 50, p. 1851–1865

Behre, C. H., Jr., 1932, The Weston Pass mining district, Lake and Park counties, Colorado: Colo. Sci. Soc. Proc., v. 13, p. 53–75

—— 1939, Preliminary geologic report on the west slope of the Mosquito Range in the vicinity of Leadville, Colorado: Colo. Sci. Soc. Proc., v. 14, p. 49–79

Brainerd, A. E., Baldwin, H. L., Jr., and Keyte, I. A., 1930, Stratigraphic sections in southern Rocky Mountains of Colorado: Kans. Geol. Soc. Guidebook, 4th Ann. Field Conf., p. 74–96

—— 1933, Pre-Pennsylvanian stratigraphy of the Front Range in Colorado: Am. Assoc. Petroleum Geologists Bull., v. 17, p. 375-396

Brill, K. G., Jr., 1952, Stratigraphy in the Permo-Pennsylvanian zeugogeosyncline of Colorado and northern New Mexico: Geol. Soc. America Bull., v. 63, p. 809–888

Burbank, W. S., 1930, Revision of geologic structure and stratigraphy in the Ouray district of Colorado, and its bearing on ore deposition: Colo. Sci. Soc. Proc., v. 12, p. 151–192

—— 1932, Geology and ore deposits of the Bonanza mining district, Colorado: U. S. Geol. Survey Prof. Paper 169, 166p.

Burbank, W. S., and Goddard, E. N., 1937, Thrusting in Huerfano Park, Colorado, and related problems of orogeny in the Sangre de Cristo Mountains: Geol. Soc. America Bull., v. 48, p. 931–976

Butler, C. R., 1949, Geology of the northern part of the Sangre de Cristo Mountains, Colorado (Abstract): Geol. Soc. America Bull., v. 60, p. 1959–1960

Crawford, R. D., 1913, Geology and ore deposits of the Monarch and Tomichi districts, Colorado: Colo. Geol. Survey Bull. 4, 317 p.

Cross, W., 1894, Description of the Pikes Peak sheet: U. S. Geol. Survey Folio 7

—— 1910, Description of the Engineer Mountain quadrangle, Colorado: U. S. Geol Survey Folio 171

Cross, W., and Ransome, F. L., 1905, Description of the Rico quadrangle, Colorado: U. S. Geol. Survey Folio 130

Cross, W., Howe, E., Irving, J. D., and Emmons, W. H., 1905, Description of the Needle Mountains quadrangle, Colorado: U. S. Geol. Survey Folio 131

Donner, H. F., 1949, Geology of the McCoy area, Eagle and Routt counties, Colorado: Geol. Soc. America Bull., v. 60, p. 1215–1248

Eldridge, G. H., 1894, Description of the sedimentary formations, Anthracite and Crested Butte quadrangles, Colorado: U. S. Geol. Survey Folio 9

Emmons, S. F., 1882, Geology and mining industry of Leadville, Lake Co., Colorado: U. S. Geol. Survey Ann. Report 2, p. 215–230

Emmons, S. F., Irving, J. H., and Loughlin, C. F., 1927, Geology and ore deposits of the Leadville district: U. S. Geol. Survey Prof. Paper 148, 368 p.

Endlich, F. M., 1874, Geology of Section B, San Luis division: 7th Ann. Rept., U. S. Geol. Geog. Survey Terr. (Hayden), p. 323–334

—— 1877, Report on southern Colorado: 9th Ann. Rept., U. S. Geol. Geog. Survey Terr. (Hayden), p. 103–235

Flower, R. H., 1952, Cephalopods from the Harding and Manitou formations of Colorado: Jour. Paleontology, v. 26, p. 505–518

Gabelman, J. W., 1952, Structure and origin of northern Sangre de Cristo Range, Colorado: Am. Assoc. Petroleum Geologists Bull., v. 36, p. 1547–1612

—— 1953, Definition of a mineral belt in south central Colorado: Econ. Geology, v. 48, p. 177—210

Hills, R. C., 1899, Description of the Elmoro quadrangle, Colorado: U. S. Geol. Survey Folio 58

Johnson, J. H., 1929, Contribution to the geology of

the Sangre de Cristo Mountains of Colorado: Colo. Sci. Soc. Proc., v. 12, p. 1–21

Johnson, J. H., 1934, Paleozoic formations of the Mosquito Range, Colorado: U. S. Geol. Survey Prof. Paper 185-B, p. 15–43

—— 1944, Paleozoic stratigraphy of the Sawatch Range, Colorado: Geol. Soc. America Bull., v. 55, p. 303–378

—— 1945, A resume of the Paleozoic stratigraphy of Colorado: Colorado School Mines Quart., v. 40, p. 1–109

Kirk, E., 1930, The Harding sandstone of Colorado: Am. Jour. Sci., 5th ser., v. 20, p. 456–467

—— 1931, The Devonian of Colorado: Am. Jour. Sci., 5th ser., v. 22, p. 222–240

Koschman, A. H., and Wells, F. G., 1946, Preliminary report on the Kokomo mining district, Colorado: Colo. Sci. Soc. Proc., v. 15, p. 51–112

Langenheim, R. L., Jr., 1952, Pennsylvanian and Permian stratigraphy in Crested Butte quadrangle, Gunnison County, Colorado: Am. Assoc. Petroleum Geologists Bull., v. 36, p. 543–574

Lovering, T. S., and Tweto, O. L., 1944, Preliminary report on geology and ore deposits of the Minturn quadrangle, Colorado: U. S. Geol. Survey Open-File Report

Maher, J. C., 1950, Pre-Pennsylvanian rocks along the Front Range: U. S. Geol. Survey Prelim. Chart 39, Oil and Gas Investig. Ser.

—— 1953, Permian and Pennsylvanian rocks of southeastern Colorado: Am. Assoc. Petroleum Geologists Bull., v. 37, p. 913–939

McKee, E. D., 1951, Sedimentary basins of Arizona and adjoining areas: Geol. Soc. America Bull., v. 62, p. 481–506

Melton, F. A., 1925, Ancestral Rocky Mountains of Colorado and New Mexico: Jour. Geology, v. 33, p. 84-90

Siebenthal, C. E., 1910, Geology and water resources of the San Luis Valley, Colorado: U. S. Geol. Survey Water Supply Paper 240, 128 p.

Singewald, Q. D., and Butler, B. S., 1932, Suggestions for prospecting in the Alma district, Colorado: Colo. Sci. Soc. Proc., v. 13, p. 89–131

Stark, J. T., *et al.*, 1949, Geology and origin of South Park, Colorado: Geol. Soc. America Mem. 33, 188 p.

Stone, J. B., 1934, Limonite deposits at the Orient mine, Colorado: Econ. Geology, v. 29, 317–329

Sweet, W. C., 1954, Harding and Fremont formations Colorado: Am. Assoc. Petroleum Geologists Bull., v. 38, p. 284–305

Tweto, Ogden, 1949, Stratigraphy of the Pando area, Eagle County, Colorado: Colo. Sci. Soc. Proc., v. 15, p. 149–235

U. S. Geological Survey, 1935, Geologic map of Colorado: Compiled by W. S. Burbank *et al.*; edited by G. W. Stose; scale 1:500,000

Vanderwilt, J. W., and Fuller, H. C., 1935, Correlation of Colorado Yule marble and other Paleozoic formations on Yule Creek, Gunnison County, Colorado: Colo. Sci. Soc. Proc., v. 13, p. 439–463

Walcott, C. D., 1892, Preliminary notes on the discovery of a vertebrate fauna in Silurian (Ordovician) strata: Geol. Soc. America Bull., v. 3, p. 153–172

Worcester, P. G., 1919, Molybdenum deposits of Colorado: Colo. Geol. Survey Bull. 14, 131 p.

THE CALIFORNIA COMPANY, HARVEY, LOUISIANA

MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, APRIL 28, 1955

Appendix: Description of Measured Sections

Section 1

Measured along the crest of the range southward from the Precambrian contact beginning about 400 yards south of Galena Peak in Secs. 21 and 27, T. 47 N., R. 10 E.

Only the basal portion of the Leadville limestone could be measured here because of structural complications.

	Feet
Leadville limestone (Mississippian):	
54. Limestone, dark gray, weathers medium to dark gray, fine-grained, contains large dark-gray chert nodules.	30.0
53. Limestone, medium gray, weathers light gray, fine-grained.	13.5
52. Limestone, dark gray, fine-grained, sandy.	4.5
51. Limestone, medium light gray, weathers light gray, sandy zones interbedded with zones that contain no sand.	8.0
50. Limestone, medium dark gray, fine-grained.	34.5
49. Limestone, medium light gray, weathers very light gray, fine-grained.	4.5
48. Limestone, medium gray, massive, shattered appearance.	14.0
47. Limestone, medium light gray, weathers same, massive, fine-grained, shattered appearance, small calcite veinlets.	16.5
Partial thickness of Leadville limestone.	125.5

Disconformity. In places there is a little thin-bedded material that appears to be weathered Chaffee formation.

	Feet
Chaffee formation (Upper Devonian):	
Dyer dolomite member:	
46. Dolomite, medium light to light gray, weathers grayish yellow, fine-grained, does not crop out plainly but weathers into slabs a few inches across.	15.0
45. Dolomite, light gray, weathers yellowish gray, contains small angular fragments of chert that weather out forming a very rough surface, very hard.	14.0
44. Dolomite, light gray, weathers very light gray to grayish yellow, fine-grained, conchoidal fracture.	7.5
43. Dolomite, light gray to grayish orange, weathers yellowish gray. Much of this zone contains chert in wirelike stringers that penetrate the dolomite in all directions. The chert stringers weather out into ridges up to half an inch high, forming a rough surface. Where there are no chert stringers the rock breaks with a conchoidal fracture.	25.0
42. Dolomite, light gray, weathers yellowish gray, fine-grained, contains a little chert.	1.5
41. Dolomite, light gray, weathers very light gray, fine-grained, rough weathered surface.	8.5
40. Dolomite, pinkish gray, weathers grayish yellow, fine-grained, distinctly bedded.	9.0
39. Dolomite, light gray, weathers yellowish gray, fine-grained, subconchoidal fracture, weathered surface has sandpaperlike texture.	6.0
38. Dolomite, light gray, weathers yellowish gray, fine-grained, subconchoidal fracture, contains small sandy laminae that weather out as fine lines in the direction of bedding.	9.0
37. Dolomite, light gray, weathers yellowish gray, fine-grained, subconchoidal fracture.	4.5
36. Dolomite, sandy, and dolomitic sandstone, light olive gray, weathers dark yellowish orange, breaks easily into small fragments—a transition zone between the Parting and Dyer members of the Chaffee formation.	7.0
Total thickness of Dyer dolomite member.	107.0

Contact between the Parting and Dyer members appears gradational.

Parting quartzite member:	
35. Quartzite, conglomeratic, most of coarse sand size, but with grains up to a quarter of an inch in diameter.	3.0
34. Quartzite, light gray, medium- to fine-grained, very resistant.	8.5
33. Conglomerate, quartzite, and dolomitic sandstone. The conglomerate contains pebbles of (Fremont) dolomite up to 6 inches in diameter; the matrix is coarse quartz sand. The dolomitic sandstone is light olive gray and weathers dark yellowish orange. The thickness of this unit ranges from 4 to 10 feet.	7.0
Total thickness of Parting quartzite member	18.5
Total thickness of Chaffee formation	125.5

Disconformity. Most prominent one in the section.

	Feet
Fremont dolomite (Upper Ordovician):	
32. Covered. Probably weathered zone at top of the Fremont.	5.0

	Feet
31. Dolomite, light gray, fine-grained.	3.5
30. Dolomite, weathers medium light gray, thick-bedded, sandy.	16.0
29. Dolomite, light gray and yellowish gray in a splotched pattern, sandy.	6.5
28. Dolomite, weathers medium light gray. This zone is poorly exposed. There were a few crinoid stems near the top, and a sandy zone near the middle.	114.5
27. Dolomite, medium light gray, weathers yellowish gray, fine-grained, distinctly bedded, beds average about 6 inches thick.	30.0
26. Dolomite, medium gray, weathers medium light gray. The middle portion is thick-bedded, but the upper and lower parts are medium-bedded.	40.0
25. Dolomite, medium gray, massive, slightly fossiliferous (few horn corals), many small calcite veinlets.	63.0
24. Sandstone, dolomitic, dark weathered surface.	4.0
Total thickness of the Fremont dolomite	282.5

Disconformity.

Harding sandstone (Middle Ordovician):	Feet
23. Quartzite, light olive to dark olive, thin-bedded.	10.0
22. Sandstone and shaly sandstone interbedded, dark gray to dark brown.	3.5
21. Quartzite, medium gray, massive, top part thick-bedded and lighter in color.	35.0
20. Quartzite and shale interbedded. The quartzite is thin-bedded and fine-grained. The shale is dark gray.	17.0
19. Quartzite, medium light gray, massive, fine-grained.	25.0
18. Shale, dark gray to black, lenses of fine sandstone.	1.5
17. Quartzite, nearly white, fine-grained.	5.0
16. Shale (or siltstone), olive to dark yellowish orange, seems to vary in thickness laterally.	14.0
Total thickness of Harding sandstone	111.0

Disconformity.

Manitou dolomite (Lower Ordovician):	Feet
15. Dolomite, medium gray, top part sandy with occasional fine sandstone lenses.	8.5
14. Dolomite, light to medium light gray, weathers yellowish gray, fine-grained.	1.5
13. Dolomite, medium gray, sandy, dark wirelike chert on some bedding surfaces.	3.0
12. Dolomite, medium gray, weathers light gray, silty, prominent bedding, small amount of chert.	9.0
11. Dolomite, medium gray, weathers medium light gray, thick-bedded, middle portion contains chert layers, thin stringers of chert in remainder.	25.5
10. Dolomite, medium gray, weathers light gray, minor amount of chert, rough weathered surface with sharp nodes.	2.5
9. Dolomite, medium light gray, sandy.	4.5
8. Dolomite, medium light gray, weathers yellowish gray, silty, fine-grained, weathers with long parallel ridges on bedding surface like ripple marks, but only half an inch to 1 inch from crest to crest, the ridges are sharp.	7.5
7. Dolomite, medium gray, weathers medium light gray, sandy, contains stringers of chert.	12.5
6. Dolomite, medium gray, weathers medium light to light gray, contains some layers of chert.	22.0
5. Dolomite, medium light gray, weathers medium light to light gray, silty, contains layers of chert.	8.5
4. Dolomite, weathers yellowish gray, contains fine laminae of fine sand.	4.0
3. Dolomite, interbedded layers of lighter and darker weathering dolomite. The darker layers weather to medium light gray and contain layers of chert 1–3 inches thick.	37.0
2. Dolomite, medium gray, weathers light gray, massive.	33.5
1. Dolomite, light gray, weathers yellowish gray, silty, sandy at base, contains layers of white to pink chert.	13.5
Total thickness of Manitou dolomite	193.0

Unconformity.

Precambrian: The top of the Precambrian is a brown weathered zone up to 8 feet thick. The surface of deposition was relatively smooth as relief in this exposure does not exceed 1 foot.

Section 2

Measured along the north side of South Piney Creek in Sec. 2, T. 46 N., R. 10 E. Section rather poorly exposed.

Minturn formation (Pennsylvanian and Permian (?)): The basal portion is for the most part brown quartzose sandstone.

Unconformity.

Leadville limestone (Mississippian):

9. Covered. contact accurate to 10 feet. Interval has limestone float that is

	Feet
medium gray and weathers light gray, also limestone conglomerate.	112.0
8. Limestone, medium gray, weathers medium light gray, massive, contains black chert, some limestone conglomerate.	22.5
7. Limestone, medium gray, weathers very light gray, laminated, some limestone conglomerate.	3.5
6. Limestone, medium dark gray, weathers medium light gray, massive, well-developed joint, contains black chert nodules, also fragments of black limestone. Forms highest cliffs.	17.5
5. Covered. Float contains limestone, medium gray, laminated, black chert.	50.0
4. Limestone, medium dark gray, weathers yellowish brown, massive, contains irregular masses of chert.	4.0
3. Limestone, medium dark gray, weathers yellowish brown, forms ledge at top of lowest cliff.	4.0
2. Limestone, medium dark gray, weathers light gray, very massive, black chert lenses with bedded nature, dark brown laminated chert in upper 5 feet. *Spirifer* zone 6 inches thick near base. Forms lowest cliff.	40.0
1. Covered. Float contains limestone, light gray, and limestone conglomerate.	83.0
Total thickness of Leadville limestone	336.5

Disconformity (?). Contact with Chaffee not exposed, but located accurate to 10 feet.

Chaffee formation (Upper Devonian).

Short Notes

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DEVELOPMENT OF MEDIAN ELEVATIONS IN OCEAN BASINS

BY H. W. MENARD

Introduction

Recent geophysical measurements reveal that ocean basins are profoundly different from continents. The hypothesis that continents formerly existed in some places where ocean basins are now found is correspondingly less acceptable. The hypothesis of epeirogeny in ocean basins, however, is not adversely affected by the geophysical measurements; on the contrary, it is supported, although the evidence is not everywhere compelling.

Geological evidence for widespread epeirogeny in ocean basins dates back to Darwin's theory of the origin of atolls and barrier reefs. The distribution of drowned ancient islands or guyots (Hess, 1946) indicates large additional areas in the ocean basins where great changes in depth have occurred. Epeirogeny may have produced the trans-oceanic land bridges, required by paleontologists and paleogeographers, in the form of transient links (Willis, 1932). Hess (1955) has recently emphasized the role of epeirogeny in ocean basins by proposing that the many oceanic rises and ridges are produced by this mechanism. He suggests that serpentine is formed by the reversible reaction of water with olivine in the mantle, and that the resulting volume change causes epeirogeny.

The evidence and the hypotheses here proposed have been discussed with many colleagues. Comments by Robert Arthur, Walter Elsasser, E. L. Hamilton, R. R. Revelle, and Walter Munk have been particularly helpful. Pre-publication copies of Plate 120, a bathymetric chart of the Pacific Basin, from the new Times Atlas were made available through the kindness of John C. Bartholomew, Jr. This paper represents in part results of research supported by the Office of Naval Research and the National Science Foundation.

Concentration of Submarine Elevations along the Median Line of Ocean Basins

The writer has attempted to locate the median lines of all ocean basins in order to compare them with the locations of submarine elevations. The technique used is to draw a family of lines approximately parallel to and equidistant from the 1000-m contour by swinging arcs, with string and chalk, on a 50-cm diameter globe. The intersections of arcs with equal radii are taken as midpoints, and the midpoints are connected to establish median lines. Thus, not only an entire ocean but also any major convexity of the margin of an ocean has a median line. Examples of convexities are found on the west side of Africa and in the western Pacific in the great curve bounded by New Guinea, the Philippines, and Japan.

The results of this construction (Fig. 1) are shown on a Mercator projection chart along with the best-known locations of the crests of oceanic rises and ridges, and the position of the 1000-m contour used in the construction where it does not closely correspond to a continental margin. In most places the 1000-m contour can be taken, without question, from a good chart. Elsewhere, particularly along the western border of the Pacific, a choice must be made within a zone of islands and basins. The contour selected lies on the continental side of the oceanic basin nearest the continent. Oceanic basins are identified by the rule of Ewing and Press (1955, p. 2), based on earthquake surface waves, that any place deeper than 2000 fathoms is underlain by an oceanic type of crust. Even with the help of this rule, the position of New Zealand remains a puzzle. It may be regarded as a small "continent" or a continental island. It is here considered a continent, and the median lines south of 20°S. Lat. are relative to

New Zealand. North of 20° S. Lat. the median lines are relative to Australia and New Guinea although New Zealand would give almost the same position.

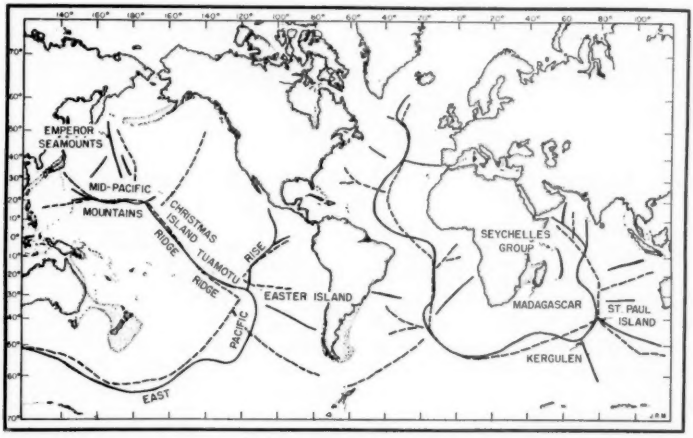

FIGURE 1.—LOCATION OF THE CRESTS OF SUBMARINE MEDIAN RIDGES AND RISES (solid line) COMPARED WITH THE GEOMETRICAL MEDIAN LINE (dashed line) OF THE OCEAN BASINS

Points on the geometrical median line are equidistant from the margins of the continents as indicated by the 1000-m contour. The 1000-m contour is shown by dotted shading where it does not almost coincide with the shore line and where it is pertinent to the construction of the geometrical median line.

Median Elevations of the Atlantic, Antarctic, Indian, and South Pacific Oceans

The median line follows the sinuous crest of the oceanic rises and ridges through the Atlantic, Indian, and Antarctic oceans with remarkable consistency. Even in the south Pacific the two lines agree in trend although they are as much as 1000 km apart. The agreement is particularly noteworthy in the south Indian Ocean where both the crest and the median line branch. Elsewhere the minor branches from the main rise have the same trends and in many places the same positions as the spurs from the central median line. Several prominent features such as the Walvis Ridge and the Newfoundland Rise are not duplicated by spurs, and several spurs are not duplicated by known rises in the south Atlantic. On the whole, however, the discrepancies merely emphasize the generally close agreement in the lines. It is at least not improbable that the agreement signifies a causative relationship—that is, the rises are in the middle because it is the middle.

An objective comparison is possible only if some well-defined line, such as the 1000-m contour, is selected as the edge of the ocean basins. However, the process that ceters then rises between continents may not be sensitive to minor bends in the continental margin, or the margins may have shifted since the process began. These uncertainties may justify some experimentation in attempting to obtain an exact fit between the median line and the crest of a rise. For example, by smoothing off relatively pointed continental projections into the Indian Ocean, such as India and Madagascar, and by removing Sumatra and Java, the median line can be superimposed on the crest of the rise and on the main branches of the rise. It would be profitless at this time to attempt to attach meaning to this particular adjustment on geological grounds.

In the southeastern Pacific the trend and position of the median line and the three spurs toward South America show a close correspondence to the known topography. The crest between New Zealand and Antarctica lies near the median line, at about 55° S. Lat.; however, the crest bends sharply to the east and is about 1000 km east of the median line from the bend to Easter Island. Submarine topography suggests that the bend may be an offset caused by faulting. The spurs of the median line are almost superimposed on the two northern branches of the rise as drawn, but the branches are not in the positions shown on most charts. The positions are based on closely spaced sounding lines taken during the Navy HIGH-

JUMP and DEEPFREEZE expeditions, and on the unpublished soundings of the Scripps SHELLBACK, CAPRICORN, and EASTROPIC and DOWNWIND expeditions. The southernmost branch of the rise and the corresponding median line are roughly parallel but widely separated. As in the Indian Ocean, it is possible to obtain an exact agreement, but the whole of South America south of the Brazilian shield must be removed—which appears unjustified unless the rise is thought to be very old.

The rises and ridges shown on the chart essentially are all parts or branches of one great median elevation. To this extent the chart supports the conclusion of Heezen (1957) that there is a world-girdling rise. However, the existence of the rise may be questioned in some places, and the position of the crest elsewhere may be poorly defined. For this reason the location of the crest as shown in Figure 1 does not agree everywhere with the location shown by Heezen.

Median Elevations of the Central and North Pacific

The median line of the central and north Pacific has four main divisions (Fig. 1): a line trending northwest from Mangareva Island to Johnston Island, and three spurs from Johnston Island trending west, northwest, and northeast. The line is followed from Mangareva to Johnston and thence along the western spur by a series of great submarine ridges which appear to be connected. The southernmost ridge is the extraordinary Tuamotu Ridge with an area of approximately 200,000 sq. km shoaler than 3 km. About 100 atolls and seamounts including guyots are distributed on the ridge. To the northwest is the Christmas Island Ridge which has an area of roughly 100,000 sq. km at or less than 3 km depth and an almost continuous line of atolls, banks, and seamounts along its crest. This ridge abuts the west-trending Marcus-Necker Rise which, in its eastern half (the Mid-Pacific Mountains), has an area of roughly 60,000 sq. km shoaler than 3 km. The Mid-Pacific Mountains includes about 50 seamounts, most of which are guyots, distributed all over the ridge. The Mid-Pacific Mountains and the Tuamotu Ridge and atolls are similar in every respect except that coral has capped the Tuamotu peaks. The three connecting ridges lying along the median line of the Pacific Basin for almost 10,000 km probably would have been named the "Mid-Pacific Ridge" if they had all been discovered at the same time. As it is, the separate names seem adequate. The northwest-trending spur lies somewhat east of, but parallel to, the Emperor Seamounts which include a number of large guyots (Dietz, 1954, p. 1209–1211). The seamounts rise above a continuous narrow ridge at least 1 km high (Bezrukov and Udincev, 1955, translation by E. R. Hope available through the courtesy of the Defense Research Board of Canada). The northeast-trending spur lies in a relatively well-explored part of the Pacific Basin, and it does not appear to be related to any continuous ridge or line of seamounts, although the eastern end of the Mid-Pacific Mountains has a conspicuous northeasterly trend and lies just north of the spur.

Differences in Morphology of Median Elevations

Median elevations of ocean basins include a variety of types which are gradational from very broad rises with gentle slopes to narrow, steep-sided ridges (Fig. 2). If these essentially continuous elevations are produced by the same process, why do different types of topography exist? They might be produced because of regional crustal differences, but seismology attests to the remarkable uniformity of the oceanic crust. Size of basins might have an effect, but the broadest rises and narrowest ridges in the Pacific are at the same distance from continents.

The various gradational forms of median elevations may be stages in development. The following sequence of forms is suggested:

(1) A very broad, seismically active rise with many volcanic islands and seamounts but few if any guyots. The southern East Pacific Rise is an example. Absence of known guyots may somewhat reflect sparse soundings, but many seamounts have been found on this rise, and none are guyots.

(2) A narrower, steeper, seismically active rise with volcanic islands and guyots side by side. The Mid-Atlantic Ridge in the North Atlantic is an example of this stage.

(3) Narrow, steep-sided, seismically inactive ridges lacking volcanic islands but with many guyots and atolls. Examples are the Tuamotu Ridge and Mid-Pacific Mountains.

It is difficult to fix in this sequence the position of the 45,000-mile-long trough that is said to exist on the crest of all oceanic rises (Ewing and Heezen, 1956, p. 80). The writer and his colleagues at Scripps Institution have not observed a continuous trough in the supposed

position in the northeastern Pacific. A special survey during the DOWNWIND Expedition did not find any apparant trough for 300 miles along the crest of the East Pacific Rise.[1] It appears that a median trough may not be characteristic of all rises and ridges but only of the intermediate form. The central trough may not yet have formed in the young stage, but what happens to it in the old stage? It may be filled by volcanic rocks. However, parallel troughs have been found in the southern section of the Mid-Atlantic Ridge (Heezen, personal communication). It is possible that the flanks of a broad rise subside on the outside of each trough and that a narrow central ridge remains.

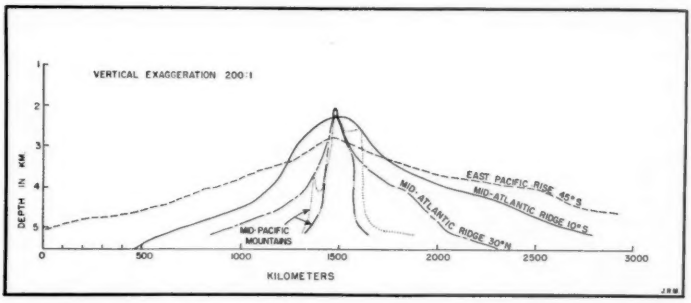

FIGURE 2.—TRANSVERSE PROFILES OF SUBMARINE MEDIAN ELEVATIONS
Showing the gradation from the broad East Pacific Rise to the steep-sided, narrow Mid-Pacific Mountains.

Permanence of Oceanic Rises

The proposed sequence of development from rises to ridges implies that oceanic rises are not permanent features. Impermanence is suggested by other lines of evidence. One of the great belts of earthquakes is centered on the crest of the world-girdling oceanic rise (Gutenberg and Richter, 1954; Ewing and Heezen, 1956, p. 78). Volcanoes abound on rises, and many have subsided to form guyots and atolls. Heat-flow measurements have been made on the East Pacific Rise, the Albatross Plateau, and the Mid-Atlantic Ridge, and values as great as twice the heat flow characteristic of continents and ocean basins are common (Bullard *et al.*, 1956; Roger Revelle, personal communication; R. P. von Herzen, personal communication). If these widely distributed measurements are typical of the intervening areas of the rise, the high values indicate an active process and instability.

The permanence of continents and ocean basins is in dispute, but there can be little doubt that each has some highly stable characteristics, namely:

(1) A distinct topographic level that is so widespread as to dominate the hypsographic curve of the earth (Kossinna, 1921, p. 36).

(2) A characteristic suite of rocks that is statistically different if not completely so (Gilluly, 1955, p. 8–10).

(3) A typical, apparently layered, crustal structure as revealed by seismic methods (Ewing and Press, 1955, p. 2–4).

A continent cannot be changed into an ocean basin without changing 35 km of one type of rock into 5 km of another type and lowering the surface by more than 5 km.

In contrast, an oceanic rise appears to have few if any stable characteristics. Rises are, on the average, 1 km shallower than ocean basins, but there is no distinct boundary between the two. Rises characteristically have the cross sections of broad arches, although in most places the arch is apparently faulted. The top is not level. Rises cover a sizable portion of the earth's surface. If any distinct intermediate level existed, it would be reflected in the hypsographic curve of the earth, but it is not.

Except for the Indian Ocean, the petrology of igneous rocks on oceanic rises is based almost wholly on samples collected subaerially

[1] Doctor Heezen has inspected all the DOWNWIND echograms in this area with the writer, in a preliminary attempt to standardize topographic terminology in the Pacific and Atlantic basins. Heezen believes that the soundings may show a small median trough and certainly do not rule out the possibility of one.

from the island peaks of large submarine volcanoes. The peaks of volcanoes are, notoriously, the locus of magmatic differentiates that are not representative of the main volume of the cone so that the petrology of most rises remains obscure. Efforts have been made to show that all oceanic rises are composed of rocks that are not typical of oceanic basins, but the efforts are based on emphasizing the atypical differentiates on the peaks at the expense of the main volume of the subaerial part of the volcanoes. Gilluly (1955, p. 9), for example, gives a table of examples of oceanic occurrences of rocks more siliceous than basalt and refers to a "rhyolite" on Easter Island citing Bandy (1937). Bandy in the cited work makes it clear that the rhyolitic obsidian is merely a differentiate of basalt, and states (p. 1607) "... the amount of obsidian on the island is 0.02 per cent of the amount of basalt that has crystallized." Hess (1954, p. 344) generalizes that basaltic vulcanism is typical of oceanic rises.

The Mid-Indian Ridge may have exceptional rocks in considerable volume. The difficulty in drawing a firm conclusion comes not in identifying the rocks but in identifying the Mid-Indian Ridge. The western half of the ocean floor is mountainous (Fairbridge, 1955, p. 161), and the Mid-Indian Ridge is not easily distinguished from the Seychelles and Maldive ridges. As drawn by Fairbridge (1954, unnumbered figure) the Mid-Indian Ridge is a normal oceanic rise capped by volcanoes. The Seychelles Islands and Maldive and Laccadive islands, however, rise from very shallow, extensive ridges; the Seychelles Islands contain granite and may have continental structure (Darwin, 1889, p. 146, 260). If the various ridges are all similar features or parts of one feature, granitic rocks occur in at least one place in an oceanic rise. However, the Seychelles Ridge appears to be a separate topographic feature perhaps a small continental outlier not unlike near-by Madagascar except that it has been truncated. The granite proves that continental rocks may be produced in ocean basins (Gilluly, 1955), but it does not prove that it is any more likely to occur on an oceanic rise than elsewhere in the basins.

The East Pacific Rise at 45° S. Lat. has a normal oceanic crust and mantle, although the second layer is unusually thick (George Shor, personal communication). This confirms the indications of surface-wave dispersion in the area (Oliver *et al.*, 1955, p. 936). The Albatross Plateau west of Central America likewise has a normal oceanic crust and mantle (R. W. Raitt, personal communication). At 10° S. lat. on the East Pacific Rise, and on the Mid-Atlantic Ridge normal mantle velocities have not been observed. Instead, the maximum velocities are 7.2 to 7.7 km/sec (Raitt, 1956, p. 1636; Ewing and Ewing, 1957). Normal oceanic crustal velocities are found at 10° S. Lat. on the East Pacific Rise, but they are lacking in some places on the Mid-Atlantic Ridge.

Abnormal velocities under rises are intriguing, but the normal crustal sections are more significant with regard to the permanence of rises. The southern part of the East Pacific Rise is 2000 km wide and 2 km high—one of the largest oceanic rises in the world—yet the crustal section is essentially the same as that under the deep basins of the oceans.

In summary, oceanic rises do not have a distinct topographic level, nor do they appear to have a characteristic suite of rocks or crustal section. Observations do not rule out the possibility that broad rises may be temporary features which are elevated and then subside, leaving narrow ridges capped with guyots and atolls.

Origin of Median Topography

Features as large as oceanic rises must be isostatically compensated. Inasmuch as some rises have a normal oceanic crustal section, the compensation occurs in the mantle. The indicated horizontal differences in the mantle have not been detected except by the fact that rises exist. Nevertheless, it appears that some process in the mantle causes localized density changes adequate to produce epeirogenic uplift of the sea floor by 2 km.

Ewing and Ewing (1957) suggest that the low maximum seismic velocities found under the Mid-Atlantic Ridge are caused by a physical mixing of mantle and crustal material. This process does not explain the relief of rises where the crustal section is of normal oceanic type.

Hess (1954, p. 346–347) proposes that serpentinization of the mantle is a common cause of oceanic epeirogeny. The reaction has the virtue of occurring wholly within the mantle and being reversible, and because it is exothermic it also offers an explanation for the high heat flow associated with rises.

The process that centers rises in ocean basins has the following characteristics:

(1) It is sensitive to the margins of the basins.

(2) It is capable of acting at distances as great as half the width of the Pacific Basin—more than 5000 km or 50° of arc.

(3) It acts on both sides of a basin at one time.

(4) It appears to be ephemeral or intermittent.

One possible process is the convection current. If so, the high heat flow along seismically active rises implies rising currents in the middle (Bullard *et al.*, 1956, p. 177). Continuity requires sinking currents near the margins of the basins with horizontal movement toward the margins under the crust. This is the direction of movement advocated by proponents of the convection origin of island arcs and Pacific fracture zones.

Corollaries

The hypothesis of epeirogenic uplift and subsidence of median oceanic elevations has several corollaries which can be used to test its validity and the validity of other hypotheses. Some of the most important ones are briefly discussed.

Faunal migration. Although not continental, if the median ridges of the Pacific were once elevated and were contemporary they would have produced many of the effects normally attributed to continents. Individual ridges are less than 1 km deeper than many of the guyots that rise above them, so that, when the guyots were islands, the ridges were deep banks, each 1000–2000 miles long. Whether the ridges were contemporary is unknown, but the distribution of existing rises suggests that single median elevations the length of an ocean basin are normal whereas shorter detached elevations are not. The only indication of the age of the ridges is the Late Cretaceous dating of the final truncation of the guyots in the Mid-Pacific Mountains (Hamilton, 1956, p. 42). Assuming approximate contemporaneity, the median line of the Pacific Basin was marked in Mesozoic time by an almost continuous deep bank capped by numerous closely spaced islands and shallow banks. The general picture closely resembles the isthmian links of Willis (1932) and is no more than a geographical extension of the dense island chains of the Mid-Pacific Mountains (Hamilton, 1956, p. 50–52). Island stepping stones and almost continuous shallow water may have been available for faunal migration from Japan to Easter Island or even to South America. A branching structure reached almost to Kamchatka.

Continental drift. Evidence in support of continental drift is derived from the similar shapes of the Mid-Atlantic Ridge and the adjacent shores of Africa and South America. This evidence can be dismissed as not pertinent to the hypothesis because both Africa and South America are almost circled by rises that are roughly parallel to the shore lines. If the Mid-Atlantic and Mid-Indian ridges are fragments left by the drift of South America and Australia respectively, what are the rise and ridge structures of the Pacific? If the continents have only begun to drift in Early Tertiary time, how did the Cretaceous rise of the Mid-Pacific Mountains become centered relative to the present margins of the Pacific?

References Cited

Bandy, M. C., 1937, Geology and petrology of Easter Island: Geol. Soc. America Bull., v. 48, p. 1589–1609

Bezrukov, P. L. and Udincev, G. B., 1955, The northern end of the Hawaiian suboceanic ridge (in Russian): Dok. Akad. Nauk, v. 103, p. 1077–1080. Defense Research Board of Canada, translation T226R by E. R. Hope

Bullard, E. C., Maxwell, A. E., and Revelle, R. R., 1956, Heat flow through the deep sea floor: Advances in Geophysics, v. 3, p. 153–181

Darwin, C., 1889, Structure and distribution of coral reefs, 3d ed: New York, D. Appleton and Co., 366 p.

Dietz, R. S., 1954, Marine geology of northwestern Pacific: description of Japanese bathymetric chart 6901: Geol. Soc. America Bull., v. 65, p. 1199–1224

Ewing, J., and Ewing, M., 1957, Seismic-refraction measurements in the Atlantic Ocean basins, in the Mediterranean Sea, on the Mid-Atlantic Ridge, and in the Norwegian Sea: Paper presented to the Annual Meeting of the Seismol. Soc. America

Ewing, M., and Heezen, B., 1956, Some problems of Antarctic submarine geology: Geophys. Mon. no. 1, p. 75–81

Ewing, M., and Press, F., 1955, Geophysical contrasts between continents and ocean basins, p. 1–6 *in* Poldervaart, A., Editor, Crust of the earth: Geol. Soc. America Special Paper 62, 762 p.

Fairbridge, R., 1954, Report on limits of the Indian Ocean: Indian Ocean Science Cong. Proc., Perth, p. 18–28

——1955, Some bathymetric and geotectonic features of the eastern part of the Indian Ocean: Deep-Sea Research, v. 2, p. 161–171

Gilluly, James, 1955, Geologic contrasts between continents and ocean basins, p. 7–18, *in* Poldervaart, A., Editor, Crust of the earth: Geol. Soc. America Special Paper 62, 762 p.

Gutenberg, B., and Richter, C. F., 1954, Seismicity

of the earth: Princeton, N. J., Princeton Univ. Press, 310 p.
Hamilton, E. L., 1956, Sunken islands of the Mid-Pacific Mountains: Geol. Soc. America Mem. 64, 97 p.
Heezen, B. C., 1957, Deep-sea physiographic provinces and crustal structure: Paper presented to the 38th Annual Meeting of the Am. Geophys. Union
Hess, H. H., 1946, Drowned ancient islands of the Pacific Basin: Am. Jour. Sci., v. 244, p. 772–791
——1954, Geological hypotheses and the earth's crust under the oceans: Royal Soc. London, Proc., v. 222, p. 341–348
——1955, Serpentines, orogeny, and epeirogeny, p. 391–408 *in* Poldervaart, A. *Editor*, Crust of the earth: Geol. Soc. America Special Paper 62, 762 p.
Kossinna, E., 1921, Die Tiefen des Weltmeeres: Inst. f. Meereskunde, H. 9, 70 p.
Oliver, J., Ewing, M., and Press, F., 1955, Crustal structure and surface-wave dispersion, Pt. IV: Atlantic and Pacific Ocean basins: Geol. Soc. America Bull., v. 66, p. 913–946
Raitt, R. W., 1956, Seismic-refraction studies of the Pacific Ocean Basin, Pt. I: Crustal thickness of the central equatorial Pacific: Geol. Soc. America Bull., v. 67, p. 1623–1640
Willis, B., 1932, Isthmian links: Geol. Soc. America Bull., v. 43, p. 917–952

Scripps Institution of Oceanography, University of California, La Jolla, Calif.

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PLEISTOCENE SHORE LINES IN MARYLAND

BY C. WYTHE COOKE

The U. S. Geological Survey has recently published a report by John T. Hack (1955) describing the geology of the Brandywine area of southern Maryland. As mapped by Hack this is a small (9 by 10 miles) rectangular area extending from a line through the villages of T. B. in Prince Georges County and Mattawoman in Charles County eastward to Patuxent River. Most of this area lies within Prince Georges County, with which I am familiar from having mapped the geology of the entire county (Cooke, 1951; 1952).

Hack's work leads him to conclusions that in some part resemble my own but in greater part differ from them. They are based on a very detailed study of a small area and on a hasty reconnaissance of a somewhat larger area in southern Maryland. Inasmuch as my conclusions are based on the mapping of a larger area in Maryland preceded by long-continued studies covering the entire seaboard from Maryland to Florida, I am not convinced that his conclusions are better founded than mine.

Comparison of Hack's map with mine shows a gratifying extent of agreement. His formation-boundary lines presumably are more nearly accurate than mine because they are based on much more detailed field work. Many of the differences in location are caused by the poor registration of those lines on my printed map; the boundary lines are shifted with respect to the topographic contour lines.

Hack's map shows more geologic units than mine. He has split my "Chesapeake group (undifferentiated)" into the well-known Calvert formation and an overlying sand (mentioned in my legend), for which he proposes the name North Keys sand. The North Keys presumably forms part of the Choptank formation. Uncertainty as to the identity of this sand led me to leave the group undifferentiated.

Hack also maps separately as an unnamed loam member at the top of the Brandywine formation, of Pliocene (?) age, large eroded areas of pebbly silty clay ranging in altitude from 200 to 240 feet above sea level. This loam may be Pleistocene. It may have accumulated as a mud bank off the mouth of the Potomac during the first interglacial epoch. At the beginning of this epoch, according to my interpretation, the sea stood about 270 feet higher than now and formed the Hazlehurst terrace (Cooke, 1925, p. 29; 1954, p. 204). Then the sea shore lay 2 or 3 miles north of the Brandywine area. Later, during the same interglacial epoch, sea level dropped to 215 feet and then to 170 feet (Cooke, 1952, p. 43).

Neither Hack nor I found traces of terrace deposits corresponding to the 215-foot shore line nor to the 170-foot shore line (both included in the Sunderland formation (Cooke, 1952, p. 46) in Prince Georges County) on the deeply eroded eastern side of the Brandywine area unless the deposits shown in his Figure 14 represent the 215-foot level, as they seem to do. He says (Hack, 1955, p. 15, 16) "It was the writer's impression while making the traverses, that the dip of the [Brandywine] formation toward the [Patuxent] River in the eastern part takes place in terracelike steps rather than as an even incline."

It is quite possible that the 170-foot shore line lay east of the Brandywine area, beyond the present course of the Patuxent, whose valley in this region appears to have been eroded during the second (?) glacial stage, which I suppose to have come after the time of the 170-foot shore line and before the sea rose to the 145-foot level (Cooke, 1952, p. 45). During this glacial stage, a time of lowered sea level, a vigorous young stream, working on soft, easily eroded Miocene beds, captured the headwaters of the old Patuxent River, which had previously flowed eastward from Laurel across Anne Arundel County (Cooke, 1952, p. 48).

The 215-foot shore line presumably crossed the peninsula between the Patuxent and the Potomac estuaries somewhere near the boundary between Prince Georges County and Charles County (Cooke, 1952, p. 46). This region is now so deeply dissected by the headwaters of Mattawoman Creek and Butlers Branch that few recognizable remnants of shoreline features are to be expected. However, Hack maps many small patches of dune sand, all above the 200-foot contour line and most of them crossing the 220-foot line. (The contour interval of Hack's base map is 20 feet). Some of these patches, no doubt, accumulated along the 215-foot shore line. Patches of dune sand near Brandywine and Cheltenham, which are

somewhat higher than the others, possibly mark an unrecognized shore line somewhere near 240 feet above sea level.

Remnants of the 215- and the 170-foot Sunderland shore lines are more easily recognized west of the Brandywine area, along the valley of the Potomac estuary, which was occupied by an estuary during late Sunderland time also. The scarps behind the patches of the 215-foot shore line become progressively higher northward because the upland, capped by the Brandywine formation, slopes upward to more than 300 feet above sea level near the District of Columbia line. My map (Cooke, 1951) shows several large patches of Sunderland terrace deposits between Washington and the southern end of Prince Georges County.

One of these patches is described by Hack (1955, p. 16, figs. 15, 16), who calls attention to a 15-foot scarp separating two terraces about 6 miles west of his mapped area, on the highway to Indian Head about 1.2 miles north of Piscataway Creek. According to Hack

> "... road cuts disclosed the underlying Brandywine formation for a considerable distance.... The drop in the upland surface corresponds with a drop in the base of the gravel, and in the base of the loam, and also corresponds with a marked change in lithology. The gravel north of the topographic unconformity is massive coarse grained and contains little sand. It is overlain by a thin deposits of silty loam. South of the break, cross-bedded sand is the dominant constituent of the lower part of the deposit. A layer of coarse boulders forms the base. The loam is much thicker and consists of sandy clay loam overlain by silt loam. Clearly this group of outcrops discloses an unconformity between two terrace deposits each of which is locally flat lying."

Hack clearly recognized these outcrops as terrace deposits lower than and differing in lithology from the Brandywine formation. Nevertheless he included them in the Brandywine on the supposition that it was built over a long period by the Potomac River as it migrated from side to side and intrenched itself in its flood-plain deposits, thus building terraces below the upland surface.

To some extent this plausible but specious hypothesis is true, for the Potomac is an ancient stream, and both the Brandywine formation and these later terrace deposits were formed along its course. However, the Brandywine should be restricted to pre-Pleistocene deposits, whereas the Sunderland formation is Pleistocene. The Sunderland, as defined for this region (Cooke, 1952, p. 46), accumulated during a cycle of oscillation of sea level; its deposition began with gravel carried by a running river and ended with silt laid down in the flooded Potomac estuary.

Terrace deposits along Patuxent River that I mapped as Pamlico and Wicomico formations are mapped by Hack as "Nottingham terraces 1–4." His "Nottingham 1 and 2" correspond for the most part to my Pamlico formation, whose surface in Prince Georges County includes two terraces, the Pamlico terrace (shore line 25 feet above sea level), and the Silver Bluff terrace (marine shore line 6 feet above sea level; estuarine shore lines may be a few feet higher because of higher tides there). The highest part of his "Nottingham 2" apparently is equivalent to the Talbot terrace (shore line 42 feet above sea level). His "Nottingham 3 and 4" represent the Penholoway terrace (shore line 70 feet above sea level), the Wicomico terrace (shore line at 100 feet), and the Okefenokee terrace (shore line about 145 feet). The terrace deposits corresponding to these three shore lines, together with those of the Talbot level, are included in the Wicomico formation in Prince Georges County (Cooke, 1951; 1952).

Hack (1955, p. 40) concludes that

> "The hypothesis of marine terracing of the Coastal Plain cannot at present be demonstrated on the upland of southern Maryland. On the contrary, the evidence suggests that no Pleistocene marine transgressions have occurred at altitudes above 100 feet at a date later than the deposition of the upland deposits."

He says (Hack, 1955, p. 39) that he found no evidence of terracing at 170 feet above sea level.

These statements seem somewhat surprising in view of the evidence presented in his book—namely, that he mapped dune sands, usually a shore feature, about 220 feet above sea level, and that he described terrace deposits about those levels in an area that I had already mapped as Sunderland formation. One can only conclude that he could not see the forest for the trees.

The area mapped by Hack is so deeply eroded that the preservation of shore features is a matter of chance, and their recognition is difficult. It is entirely too small to enable one to draw firm conclusions concerning a world-wide phenomenon such as marine terracing.

References Cited

Cooke, C. W., 1925, Physical geography of Georgia; the Coastal Plain: Ga. Geol. Survey Bull. 42, p. 19–54

Cooke, C. W. 1951, Geologic map of Prince Georges County and the District of Columbia: Md. Dept. of Geology, Mines and Water Resources
——1952, Sedimentary deposits of Prince Georges County and the District of Columbia: Md. Dept. of Geology, Mines and Water Resources Bull. 10, p. 1–53
——1954, Carolina bays and the shapes of eddies: U. S. Geol. Survey Prof. Paper 254, p. 195–206
Hack, J. T., 1955, Geology of the Brandywine area and origin of the upland of southern Maryland: U. S. Geol. Survey Prof. Paper 267, p. 1–41

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RIEBECKITE GRANITE IN THE WICHITA MOUNTAINS, OKLAHOMA

BY W. T. HUANG

Dark-green riebeckite-rich granite occurs in the Wichita Mountains of southwestern Oklahoma. The green granite forms irregular bodies within a large mass of reddish aegirite and riebeckite rocks and crops out chiefly in northeastern Elm Mountain in the east-central part of T. 3 N., R. 15 W. Smaller areas of similar rocks are on Quanah Mountain in the southeastern portion of T. 3 N., R. 14 W. This rock type has not been described in previous reports on the Wichita Mountains.

Megascopically one can recognize quartz, pinkish potassium feldspar, white plagioclase, green amphibole, and dark-green pyroxene, all having an average grain size of about 3 mm. Under the microscope, the granite is holocrystalline, hypautomorphic granular. Table 1 is a modal composition based on Rosiwal analyses of three thin sections.

Quartz occurs in anhedral grains mostly 1–4 mm in diameter, but ranging up to 6 mm. Tiny quartz grains within plagioclase in myrmekitelike blebs and graphic intergrowths are probably of replacement origin.

The potassium feldspar is orthoclase microperthite and has the same size range as quartz. Optically it appears to be monoclinic ($\perp$ X, Z$\wedge$a = 4°; $\perp$ Z, X$\wedge$b = 0°); 2V (−) = 54°, corresponding to $Or_{74}(Ab\text{-}An)_{26}$ (MacKenzie and Smith, 1956, p. 406). Numerous specks of clay and some disseminated sericite makes the potassium feldspar turbid. The plagioclase is unzoned oligoclase (An_{28}) concentrated at the margins of perthite grains.

There are two amphiboles in the rock: riebeckite and hornblende. Strongly pleochoric riebeckite with X = blue, Y = yellow green, and Z = light yellow forms stubby prisms about 2 mm long and sheaves of very thin needles 0.2–0.5 mm long. The aggregates are clustered in areas where they are associated with biotite, aegirite, iron ores, and quartz. The riebeckite has X $\parallel$ c, and low birefringence, with β about 1.688.

One of the two alteration products of riebeckite is a fine-grained yellowish pseudomorph with higher birefringence but the same parallel extinction; it partly or completely replaces some sheaves of riebeckite needles and may be acmite. A second alteration product is a mixture of magnetite and hematite with zircon. In one thin section, however, magnetite and hematite occur together as uniformly distributed interstitial grains up to 1.5 mm in diameter and make up 6 per cent of the rock.

Hornblende occurs in anhedral grains up to 4 mm across. The striking pleochroism (X = blue, Z = green or brown) suggests that it is rich in sodium. The hornblende occurs in felted aggregates and irregular crystals with clots of magnetite and ilmenite, suggesting alteration after pyroxene. In it, β ranges from 1.659 (blue variety) to 1.679.

Aegirite or aegiritic pyroxene forms euhedral to subhedral grains 0.2 mm across associated with hornblende and riebeckite. The mineral is strongly pleochroic with X = grass green and Z = yellowish green; 2V(+) = 62°. Some aegirite is partly altered to hornblende mixed with iron ores.

In the aegirite of one specimen strongly pleochroic astrophyllite occurs as an irregular grain 0.1 by 0.4 mm. The color ranges from red to yellow. Many small dark-brown fibrous aggregates associated with hornblende may also be astrophyllite. Small red granules, probably of the same mineral, are associated with clusters of riebeckite needles.

Zircon is common in all the thin sections examined; a few large euhedral crystals up to 6 mm long are zoned, similar to those found in pegmatite that cuts the Quanah granite (Larsen *et al.*, 1953). In places large zircon crystals occur together as interstitial grains. Two generations of zircon can be recognized: the large euhedral zircon crystals are primary, whereas the minute ones are secondary. The decrease in grain size from the primary to secondary is extremely abrupt. Of particular interest is the close relationship between secondary zircon and riebeckite. The zircon occurs only with altered riebeckite. The granitic rocks of the Wichita Mountains are notable for the occurrence of abundant zircon crystals, and zirconia content of the rocks must be high. Rockallite analyzed by Washington (1914, p. 297) contained 1.17 per cent zirconia; since zircon is rare in this rock, Washington concluded that zirconia is present in the acmite, and he calculated that it contains 2.6 per cent

zirconia (1914, p. 300). He suggested that the distinction between the yellow-brown acmite and the green aegirite may be due to zirconia or oxides of rare earths in the former and their absence from the latter (1914, p. 301; Washington and Merwin, 1927, p. 233, 248). A comparable view was expressed by Mathews and Watson (1953, p. 445–446) regarding alkali rhyolite dikes in northern British Columbia. A similar explanation might fit the Wichita rocks.

TABLE 1.—MINERAL COMPOSITION OF DARK-GREEN RIEBECKITE GRANITE, WICHITA MOUNTAINS
(Volume per cent)

Quartz	20
Potassium feldspar	25
Plagioclase (An_{25})	6
Riebeckite	25
Hornblende	5
Aegirite	12
Biotite	1
Muscovite	1
Zircon	1
Other accessories	4

Iron-rich olivine was found in 2 of the 20 thin sections. The mineral occurs in irregular grains or broken euhedra, 0.3–0.5 mm across, associated with clusters of riebeckite needles. Many olivine grains were selectively replaced by iron ore, a common feature of olivine alteration in the Wichita Mountains igneous rocks (Huang, 1954, p. 554–557). $2V_x$ was determined to be 58° (universal-stage measurement of six grains). According to Tröger (1956, p. 37), this corresponds approximately to Fa_{84}, a ferrohortonolite.

In one thin section radiating prehnite sheaves, 1 to 2 mm long, make up almost 2 per cent of the rock. Wavy extinction is common. Pleochroism is from pale yellow to colorless. The prehnite here recorded is like the prehnitization of gabbro in the Wichita Mountains (Huang, in press).

Minor accessories—apatite, sphene, biotite, muscovite, epidote, and magnetite—are distributed throughout the rock in euhedral to subhedral crystals.

The dark-green riebeckite-rich granite was located in the field because of a specimen presented by the late Professor J. Willis Stovall, of the University of Oklahoma, who gave preliminary information on its occurrence.

References Cited

Huang, W. T., 1954, Petrography of the troctolite of the Wichita Mountains, Oklahoma: Am. Mineralogist, v. 39, p. 549–565

——(in press), Prehnitization of gabbro: Texas Jour. Sci.

Larsen Jr., E. S., Waring, C. L., and Berman, J., 1953, Zoned zircon from Oklahoma: Am. Mineralogist, v. 38, p. 1118–1125

MacKenzie, W. S., and Smith, J. V., 1956, The alkali feldspars, III, An optical and x-ray study of high-temperature feldspars: Am. Mineralogist, v. 41, p. 405–427

Mathews, W. H., and Watson, K. D., 1953, Spherulitic alkali rhyolite dikes in the Atsutla Range, Northern British Columbia: Am. Mineralogist, v. 38, p. 432–447

Tröger, W. E., 1956, Optische Bestimmung der Gesteinsbildenden Minerale-Teil I, Bestimmungstabellen: E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart

Washington, H. S., 1914, The composition of rockallite: Geol. Soc. London Quart. Jour., v. 70, p. 294–301

Washington, H. S., and Merwin, H. E., 1927, The acmitic pyroxenes: Am. Mineralogist, v. 12, p. 233–252

DEPARTMENT OF GEOLOGY, BAYLOR UNIVERSITY
WACO, TEXAS

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PLATE 1.—PHOTOMICROGRAPHS

FIGURE 1.—Riebeckite granite from Elm Mountain, NE 1/4 Sec. 24, T. 3 N., R. 14 W. Euhedral and altered riebeckite crystals occupy most of the photograph. The rest is potassium feldspar, plagioclase, quartz, and biotite. Note that zircon occurs only with altered riebeckite (lower central and right). Others are bubbles (upper). Plain light, × 36.

FIGURE 2.—Enlargement of the altered riebeckite in Figure 1. The riebeckite is a mixture of magnetite, hematite with closely associated zircon (high relief). Plain light, × 245.

FIGURE 1

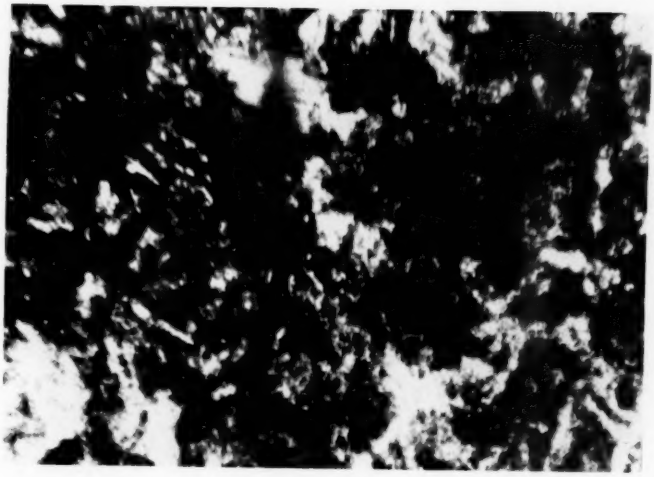

FIGURE 2

PHOTOMICROGRAPHS

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PHYSICAL CONTROL AND AGE OF DIKE FORMATION IN THE MAINE COASTAL REGION

BY CARLETON A. CHAPMAN AND PAUL S. WINGARD

About 2000 basic dikes have been studied in the Maine coastal region between Penobscot Bay and Gouldsboro Bay. The area investigated extends 45 miles along the coast and averages about 15 miles in width. For convenience it may be subdivided into three sections: (1) the Castine-Blue Hill Peninsula on the west, mapped by Smith, Bastin, and Brown (1907); (2) Mount Desert Island in the center, mapped by Shaler (1889) and Chadwick (1939; 1944); and (3) the Gouldsboro Peninsula on the east. The seven major rock units cut by the dikes in this region are correlated in Table 1.

The dikes are irregularly distributed within individual rock units as well as from one unit to another. Some have been metamorphosed to amphibolitic rocks, and many have been altered hydrothermally, but originally most were basaltic. Most of the dikes are less than 10 feet wide, and relatively few exceed a width of 60 feet. Simple dikes predominate, but many multiple dikes (with chilled internal contacts) have been found.

The following relations were first worked out in the Mount Desert Island section and subsequently were found to apply generally to the other two sections. Very simply Mount Desert Island consists of a large core of igneous rocks fringed by three older stratified units (Table 1). The core is composed of a large stock of hornblende granite, a smaller stock of biotite granite, an elongate mass of the granite of Southwest Harbor, and a gently dipping sill-like body of diorite.

The basic dikes may be subdivided into three groups on the basis of average trend: group I = N. 45°–75°E., group II = N. 0°–30°E., and group III = N. 10°–30°W. Group I, the oldest, has been metamorphosed and cuts only pre-Bar Harbor rocks; group II, next in age, cuts all units except the two younger granites; group III cuts all rock units.

Dikes of group III are about 20 times as abundant in the hornblende granite as in the biotite granite (see Table 1). Furthermore, abundant dikes in the hornblende granite, near its contact with the biotite granite, do not transect the younger biotite granite. These relations indicate that group III represents dikes of two distinct ages—one pre-biotite granite, the other post-biotite granite. These subdivisions, though difficult to distinguish in the field, may be designated as IIIa (older) and IIIb (younger). It seems well established, therefore, on the basis of trend, age of host rock, and mutual relations, that at least four distinct dike sets are represented.

Field relations indicate that room for the basic magma was provided by dilation of the walls. For simplicity this dilation is considered to have occurred in response to tensional stress acting normal to the walls. The cause of this tension is not known. The similarity of composition and widespread distribution of dikes of each set suggest that basic material was probably not derived from local pockets but from some deep, more universal layer. It is assumed, therefore, that basic melt was available throughout the area in sufficient quantities during each period of dike formation. Inasmuch as the average dike width for each set appears to be approximately the same, dike frequency is a practical measure of the degree of dilation involved. In general the amount of dilation or dike frequency should depend upon: (1) geologic age of the host rock, (2) physical conditions prevailing in the host rock, (3) physical properties of the host rock, and (4) amount and rate of deformation (tension or stretching) involved.

(1) Other things being equal, the older rock units should contain the greater number of dikes, because they should have been subjected to several periods of intrusion. Comparative study, however, shows little relation between dike abundance and age of host rock.

(2) The relevant physical conditions prevailing during dike formation include, principally, temperature and pressure. Bodies of the various rock units are so intimately associated with each other in space that for any given period of dike formation the prevailing physical conditions must have been essentially uniform.

The unmetamorphosed character of the Bar Harbor series (except near igneous contacts) indicates that the regional rock temperature during the last three periods of dike intrusion

(groups II and III) could not have exceeded a few hundred degrees Centigrade. It seems reasonable to conclude that differences in temperature and pressure prevailing during at

textures range widely. The hornblende granite, the granite of Southwest Harbor, and the diorite are hypidiomorphic granular with considerable interlocking of grains. Other

TABLE 1.—CORRELATION OF MAJOR ROCK UNITS

Castine-Blue Hill	Mount Desert Island	Gouldsboro	Age
Biotite granites	Biotite granite	Biotite granite	Mississippian (?)
----------	Hornblende granite	Hornblende granite	Mississippian (?)
----------	Granite of Southwest Harbor	Granite of Schoodic Peninsula	L. or M. Devonian
Diorite	Diorite	Diorite	L. or M. Devonian
----------	Bar Harbor series (graywacke, siltstone)	Bar Harbor series	L. or M. Devonian
----------	----------UNCONFORMITY----------		
Castine formation (pyroclastics)	Cranberry Island series (pyroclastics)	Cranberry Island series	M. or U. Silurian
----------	----------UNCONFORMITY----------		
Ellsworth schist	Bartlett's Island series (schist, gneiss)	----------	pre-Silurian

least the last three periods of dike formation were insufficient to account for differences in dilation. Physical conditions during the earliest period of dike formation may have been considerably different, but there is no evidence of this.

(3) The inherent susceptibility of a rock to fracture should depend upon its composition, texture, and structure. A rock undergoing stretching should either deform plastically or dilate by fracturing, or both. Anisotropy, expressed by foliation or bedding, should play an important role. Depending upon attitude, foliation and bedding might favor or hinder the development of dikes. Flat planar structures in rocks undergoing stretching should facilitate plastic flow and inhibit dike formation. Whereas abundant steep fractures might occur, they would tend to be limited to thin layers of rock bounded above and below by surfaces of bedding or foliation along which slippage could take place. This type of deformation would be accompanied by shearing and plastic flow, but regular through-going fractures would be less likely to form. Such structural control may explain the paucity of dikes in the gently dipping, well-foliated Bartlett's Island series (Ellsworth schist). Structural planes in the Cranberry Island series, however, are steep; many appear to have opened readily to permit development of roughly conformable sheets (dikes or sills).

Among the more nearly isotropic rocks, things equal, it seems that these rocks should be more brittle and should fracture more than the older foliated and stratified rocks.

Dikes of sets IIIa and IIIb are much more abundant in the hornblende granite than in the Bartlett's Island, Cranberry Island, and Bar Harbor series; and dikes of sets II, IIIa, and IIIb are much more abundant in the granite of Southwest Harbor than in the three older stratified series. This difference in frequency may reflect the more brittle character of the granites as compared with the more plastic, layered or stratified rock units. The frequency of dikes of sets II, IIIa, and IIIb in the diorite is considerably less dominant, however, over that of dikes in the stratified units. The apparently more plastic behavior of the diorite, as compared with the hornblende granite and granite of Southwest Harbor, may in part be due to its sill-like form and pronounced, but gently dipping banded structure.

(4) There is no evidence that the deformational rate differed appreciably during the different periods of dike formation, and it seems highly unlikely that the rate of deformation (tension or stretching) can account for the differences in degree of susceptibility to fracturing.

The amount of stretching during different periods of fracturing cannot readily be determined. Dike set I is of little help, because it is restricted to the two oldest rock units, and it is uncertain whether the low frequency of this

set is due to only a slight amount of stretching or to the relatively high plasticity of the host rock. The relative stretching involved in formation of the three younger dike sets may be more readily evaluated. A comparison of dike frequencies within host rocks predating set II (the five oldest rock units) shows set II dikes to be considerably more abundant than those of sets IIIa and IIIb combined. One may conclude, therefore, that the greater dilation accompanying set II was due largely to greater stretching during that period of dike intrusion. Furthermore, the apparent low frequency of set IIIb dikes indicates that considerably less stretching accompanied set IIIb than set IIIa dikes. In other words, for a given rock unit, dike frequency appears to be a measure of the amount of stretching.

In summary, basic dikes of at least four ages are recognized. It is clear that age and trend of the dikes are related, a fact that has aided in establishing the relative ages of certain host rocks. The average trend of successively younger dike sets has appeared to have rotated counterclockwise through nearly 90°. Except for the oldest dike set, successively younger sets represent correspondingly lower intensities of dilation and have probably resulted from successively lower intensities of stretching. Dike frequency is virtually independent of age of the host rock but varies with brittleness, attitude of planar structures, and amount of host rock stretching.

References Cited

Chadwick, G. H., 1939, Geology of Mount Desert Island, Maine: Am. Jour. Sci., v. 237, p. 355–363

——1944, The geology of Mount Desert Island (Acadia National Part): New York Acad. Sci. Trans., v. 6, p. 171–178

Shaler, N. S., 1889, The geology of the island of Mount Desert: U. S. Geol. Survey, 8th Ann. Rept., pt. 2, p. 987–1061

Smith, G. O., Bastin, E. S., and Brown, C. W., 1907, The Penobscot Bay Folio, Maine: U. S. Geologic Atlas 149

University of Illinois, Urbana, Illinois; Kansas State College, Manhattan, Kansas
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